QUARTERLY JOURNAL
OF THE
ROYAL METEOROLOGICAL SOCIETY

Editor: J. S. SAWYER

Vol. 85 OCTOBER 1959 No. 366

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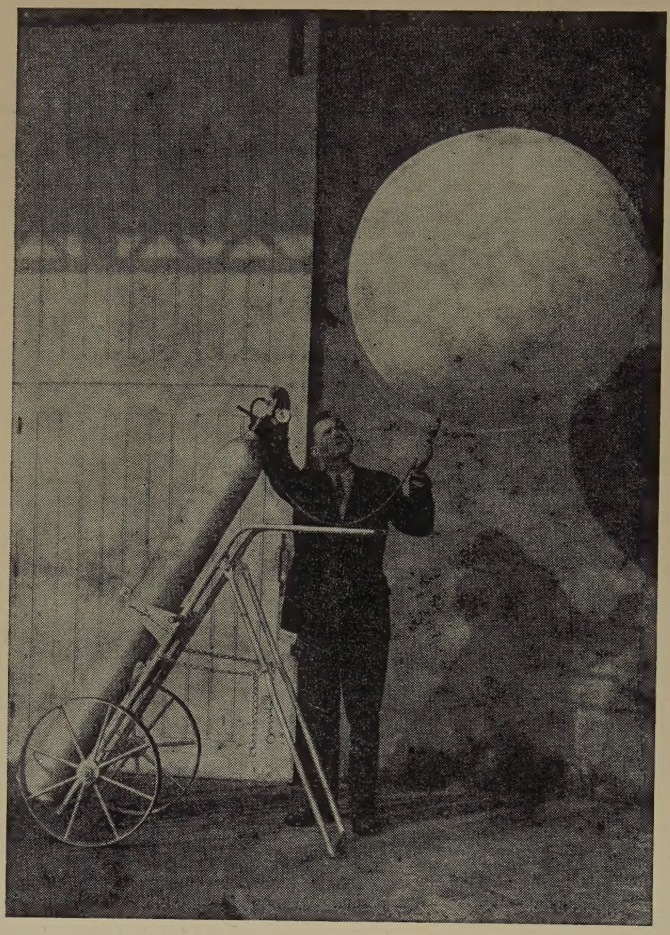

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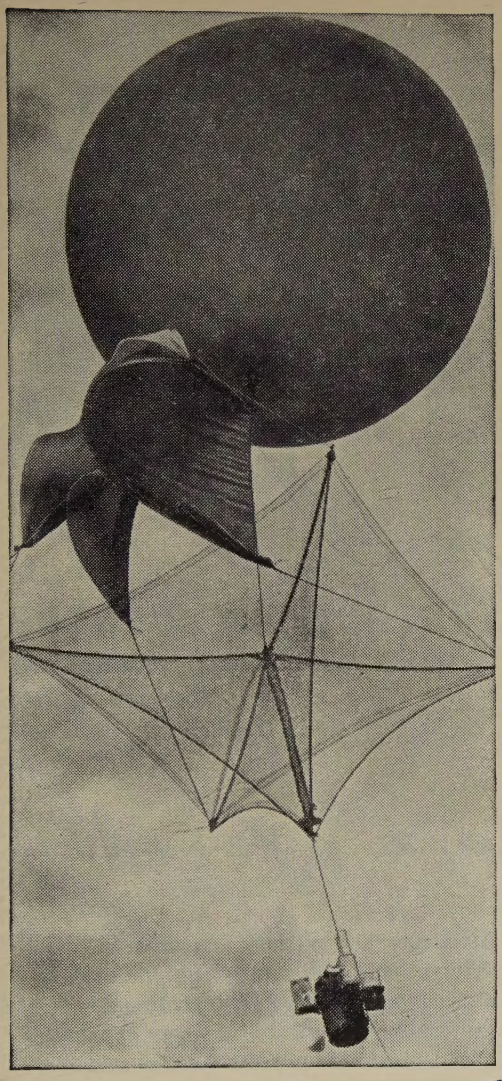

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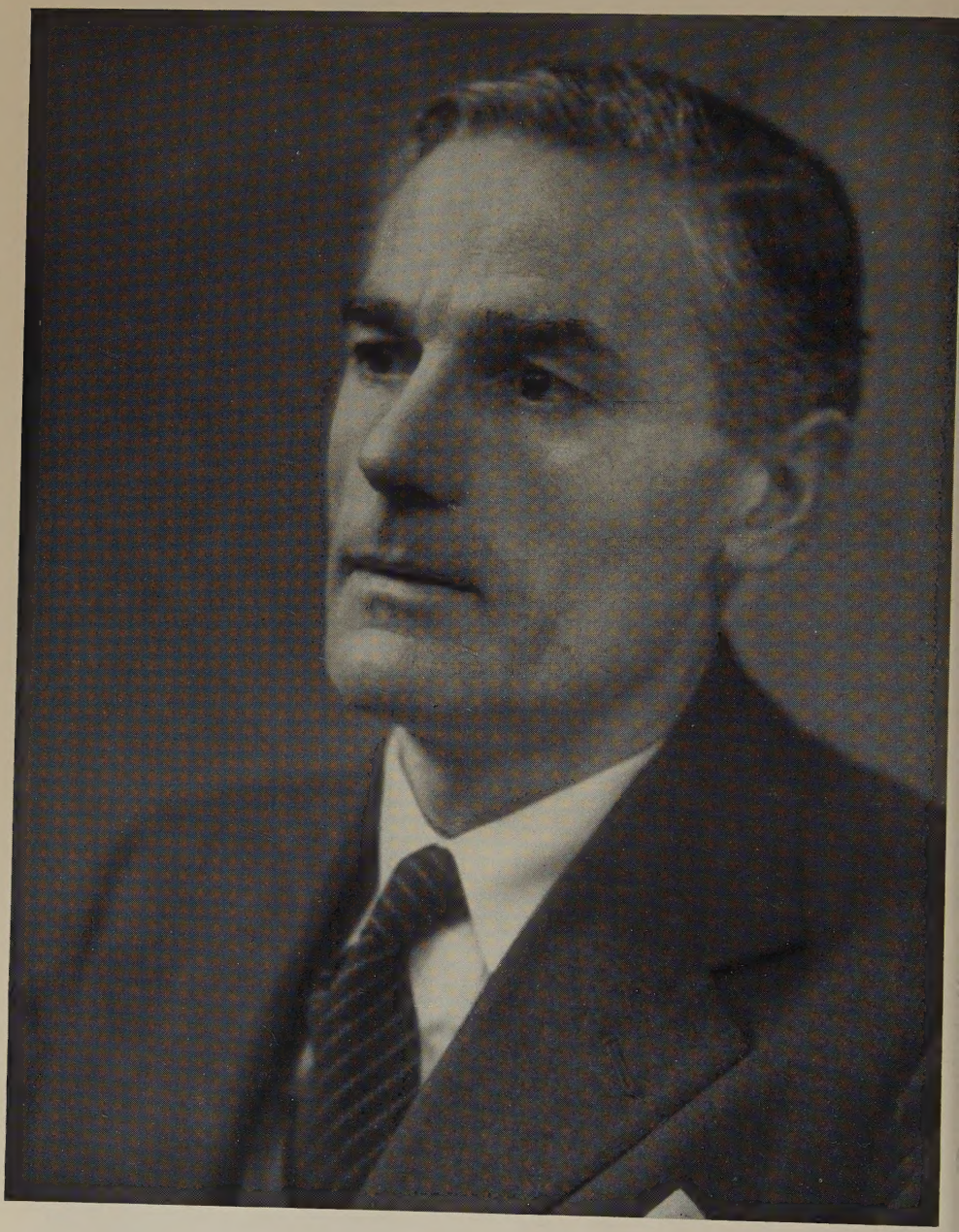

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President of the Royal Meteorological Society, 1957-1959

QUARTERLY JOURNAL
OF THE
ROYAL METEOROLOGICAL SOCIETY

Vol. 85 OCTOBER 1959 No. 366

551.510.535

The highest parts of the ionosphere

By J. A. RATCLIFFE, F.R.S.
Cavendish Laboratory, Cambridge

(Symons Memorial Lecture, delivered 1 April 1959)

In this lecture I shall describe some of the present ideas about the highest parts of the ionosphere, above about 150 km, and shall show how views about those parts have developed in the past five years or so. Although my colleagues and I are actively engaged in trying to solve some of the problems presented by this part of the ionosphere, I ought to make it clear that we have contributed only a little to what I have to tell; ideas have come from several workers, of whom Yonezawa in Japan, and Martyn in Australia deserve special mention.

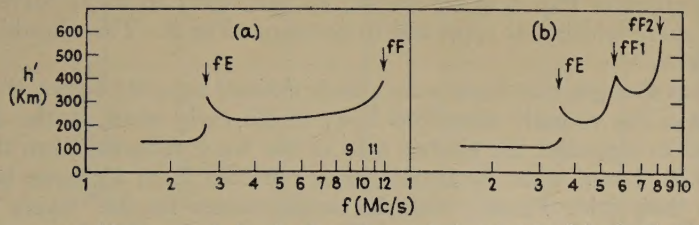

Figure 1. h′ (f) curves, or ' ionograms ' showing how the equivalent height (h′) of reflection of a radio wave depends on its frequency (f). The critical frequencies marked fE, fF, fF1 and fF2 correspond to the points marked with corresponding letters in Fig. 2.

The long-established method of exploring the ionosphere is by the study of radio-wave pulses reflected from it, and in the well-known ' ionogram ' the time of travel of these pulses, converted into an equivalent height (h′), is plotted as a function of the wave-frequency (f) in the so-called h′ (f) curve. Two examples are shown in Fig. 1. The most noticeable features are the discontinuities, or maxima, at the critical frequencies marked fE, fF1, and fF2. These frequencies correspond to maxima, or points of inflection, on an N (h) curve, of the kind shown in Fig. 2, which represents the way in which the

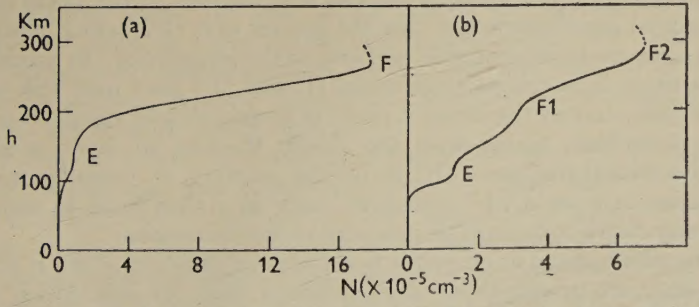

Figure 2. N (h) curves, or ' electron profiles ' to show how the electron concentration (N) varies with height (h) when the ionograms of Fig. 1 were recorded.

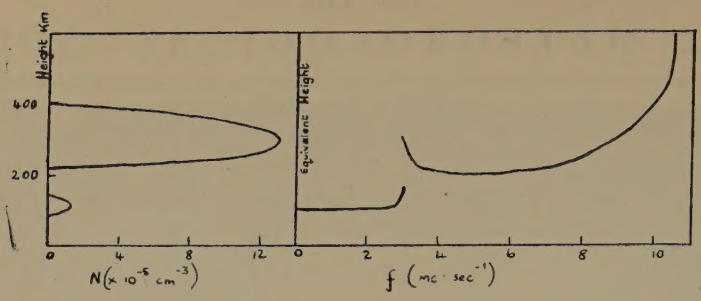

Figure 3. It was at one time thought that the ' ionogram ' on the right corresponded to the ' electron profile ' on the left, with two separate and distinct layers. It is now thought that the profile for this ionogram would be more like that of Fig. 2 (a).

electron concentration (N) varies with height (h) when the h′ (f) curves have the form shown in Fig. 1. To avoid confusion between maxima in space and in time I shall call the spatial maximum, on an N (h) curve, a ' peak,' and a point of inflection a ' ledge.' The well-marked critical frequencies on the h′ (f) curve lead to a precise knowledge of the electron concentrations at the peaks and ledges.

It was at one time thought that the critical frequencies corresponded to clearly separated layers, with well-marked minima of electron-concentration between them. An example is given in Fig. 3 which shows, on the right, an h′ (f) curve, and, on the left, the N (h) curve which was supposed to correspond to it. This should be compared with Figs. 1 and 2.

When it was thought that there were clearly defined separate layers, they were given names. Appleton has recently described how, in his early work on the ionosphere he was accustomed to represent the electric field of the wave reflected from the ionosphere by the symbol E. Later, when he found waves reflected from a greater height he used the letter F for their field. Finally, when he needed names for the ' layers ' he called the lower E and the upper F, so as to leave earlier letters of the alphabet as names for any still lower layers which might turn up, and when the F layer appeared to be doubled he named the lower part F1 and the upper part F2.

In the past few years a new method of exploration has been used, in which waves are received at the ground from rockets sent into the ionosphere. Measurements of the Doppler-shift of these waves leads to a knowledge of the electron concentration as a function of height. During the same period methods of calculation have been developed by which the same information can be derived from the h′ (f) curves. Both methods of investigation lead to the conclusion that there are *not* clearly separated layers of electrons in the ionosphere, but that the distribution is more like those of Fig. 2, deduced from h′ (f) curves, or of Fig. 4 deduced from rocket experiments.

From these newer results it might be thought that since the smaller critical frequencies fE and fF1, correspond only to weakly marked irregularities in the N (h) curve, they are of little fundamental significance, and that the greater one, fF2, which is so well-defined corresponds to some easily recognizable property of the ionosphere. In actual fact precisely the opposite is true. The critical frequencies fE and fF1 vary over the earth and with time of day, of year, and in the sunspot cycle in a smooth and regular way, which, to a first-order approximation, agrees with the simple theories of layer production. The behaviour of the critical frequency fF2 is, on the contrary, so irregular that it is usually discussed in terms of a series of ' anomalies,' and, as I shall hope to show, it does not correspond to any clearly recognizable property of the ionosphere.

The theory which seems to provide a first-order explanation of the E and F1 ' layers ' supposes that they are produced by wave-radiation from the sun which ionises one of more of the atmospheric gases, and that the resulting free electrons are afterwards lost by a recombination process at a rate proportional to the square of the electron concentration

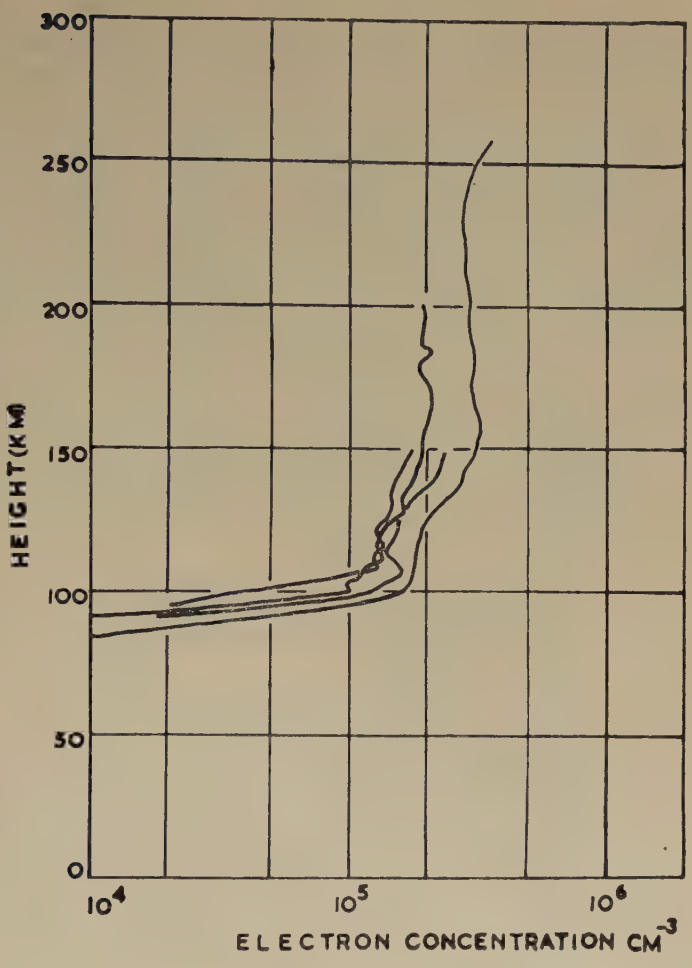

Figure 4. Some electron profiles deduced in America from measurements with rockets.

The intensities and absorption coefficients of the ionising radiations, and the recombination coefficient in the two layers, have been calculated. The intensities both increase linearly but not quite in the same way, with the sunspot number. The absorption coefficients are about what would be expected if the E-layer were formed by the absorption of soft X-rays in air, and the F1 ledge by the absorption in atomic oxygen of ultra-violet radiation in the spectral range 100 to 600 Å.

In the first-order theory the peak of the E-layer and the ledge of F1 occur because, at these two levels, there is a peak in the rate of production. The height of this peak is determined jointly by the absorbability of the ionising radiation, and by the concentration of the gas to be ionised. A small absorption coefficient will produce a peak low down and a high one will produce a peak high up.

It seems that, to a first approximation, the E layer and the F1 ledge are probably accounted for in this way, and present-day work on these regions of the ionosphere is concerned, primarily, with a study of how they depart from what would be expected on this first-order theory. These deviations are full of interest for their own sake, and shed much light on ionospheric problems, but I shall not consider them here. I shall, instead, discuss the part of the ionosphere above the F1 ledge, that is the part known as the F2 region, at heights greater than about 180 km.

The peak of the F2 layer is found at various heights, between about 240 and 600 km, depending on the time, and the place, of the investigation. Although there is considerable doubt about the atmospheric density at these heights, it can be shown that for any reasonable density the absorption coefficient of the ionising radiation would have to be very great if it were to produce a peak of electron production as high as this ; indeed no combination of radiation and absorbing gas can be suggested to account for it. The first problem which confronts the investigator of the F2 layer is therefore to find a reason for its existing at all. On the theory of a simple layer, formed by a radiation with any reasonable absorption coefficient, the peak would be lower down, so that, at the heights we are considering, the electron concentration would be decreasing with increase of height, instead of increasing.

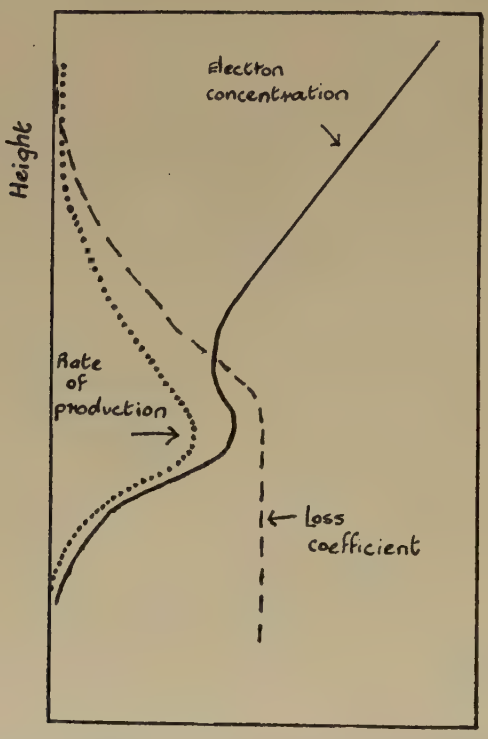

Figure 5. To illustrate Bradbury's suggestion about how the electron concentration might vary with height if the loss coefficient decreased suitably as height increased.

The theory which finds favour at present was suggested long ago (for inadequate reasons) by Bradbury, and is as follows. It is supposed that the electrons in both the F1 and F2 regions are produced by one and the same ionising radiation, which produces them most rapidly at the level of the F1 peak, near 180 km, but that the rate of loss of these electrons diminishes as the height increases. The state of affairs is illustrated in Fig. 5. If the gradient of the loss-coefficient is greater than the gradient of the density of the ionisable constituent, then it can be shown that, under equilibrium conditions, the resulting electron concentration will not show a peak but will increase steadily upwards. A mechanism of this kind will explain why the electron concentration increases upwards even at these greater heights, but it will not explain why (a) there is a peak of electrons at the level of the F1 ledge, or (b) why there is a peak in the F2 layer.

To explain the ledge at the F1 level it has been suggested that the process of electron-loss is a complex one, involving the transfer of a positive charge from an oxygen atom to a molecule, either O_2 or N_2, followed by dissociative recombination of an electron with

he molecule.* If it is further supposed that the concentration of the active molecules (O_2 or N_2) decreases upwards, then it can be shown that a process of this kind would ead at lower levels to a rate of electron-loss which was proportional to the square of the concentration of electrons and with a loss coefficient independent of height; but at high evels it would lead to a loss process in which the coefficient decreased upwards in the same way as the density of the active molecules. The rate of loss higher up would, moreover, be proportional, not to the square, but to the first power of the electron concentration. On this hypothesis the F1 peak in the distribution of electrons occurs because at that level the loss-process is independent of height. At a level only slightly higher, however, the concentration of the molecules O_2 or N_2 has decreased enough for the loss-process to become height-dependent, so that a layer of the Bradbury type can be formed, and the electron concentration increases upwards again. Whether the F1 peak is fully developed, so that there is a separate F1 layer, or whether the ' Bradbury ' increase of concentration starts so close to the peak that only a ledge is noticeable, depends on the precise values of the quantities involved. Some of the theoretical possibilities for layers in equilibrium are illustrated by the curves of Fig. 6 which show some N (h) profiles calculated from the theory mentioned above. They should be compared with the N (h) profile deduced from observations and labelled F1 and F2 in Fig. 2 (b).

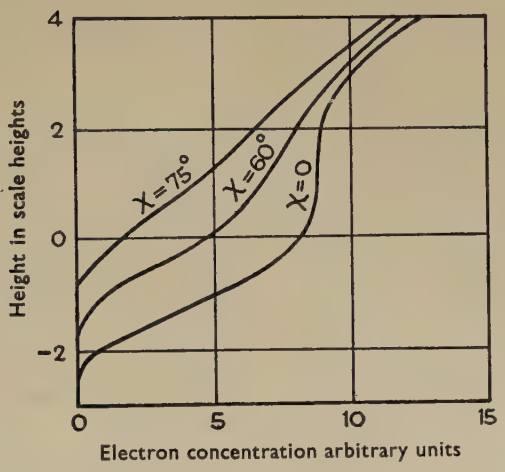

Figure 6. Some electron profiles for different angles of incidence (x) of the sun's rays calculated on the basis of a detailed theory of ' dissociative recombination ' of the kind discussed in the text. The shapes of these profiles should be compared with the experimentally determined ones shown for the F1 ledge in Figs. 2 (b) and 4.

If this hypothesis is correct the electron-loss coefficient should decrease upwards in the region above the F1 peak and it is of interest to ask whether this decrease can be demonstrated. One way of determining the loss-coefficient would be to measure the electron concentration at a fixed height throughout the night, when no new electrons were being produced, provided it could properly be supposed that there was no transport of electrons from one height to another. Unfortunately this supposition cannot be made, for reasons which will be explained later, but it is nevertheless interesting to see what results are obtained if this simple calculation is made, at different times and places. These results are shown, in terms of the ' effective attachment coefficient ' plotted against the height, in Fig. 7. There appears to be little doubt that the attachment coefficient does decrease upwards, though the scatter of the points and doubts about movements which will shortly be discussed, render the gradient of this decrease uncertain.

* The reaction involving N_2 is somewhat more complicated.

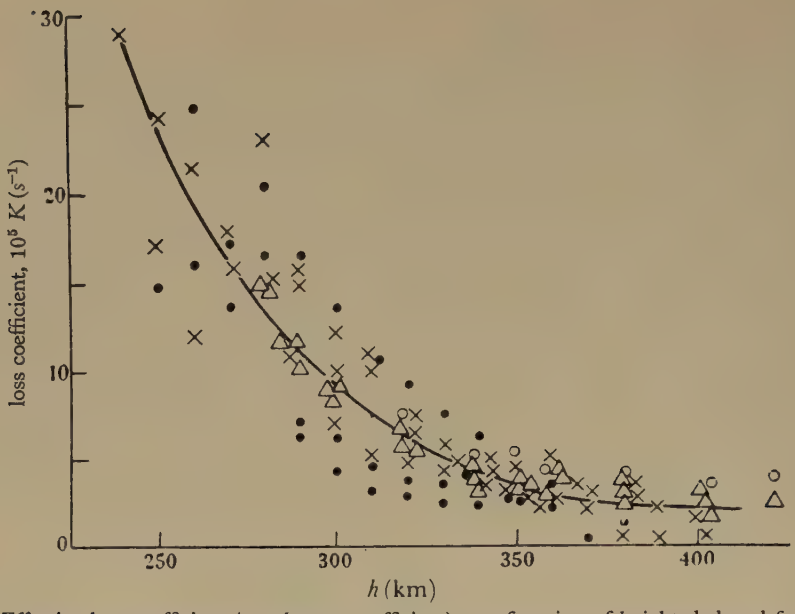

Figure 7. Effective loss coefficient (attachment coefficient) as a function of height deduced from night-tim observations.

On theoretical grounds the gradient of the loss-coefficient should be determined b the scale height of the molecule (O_2 or N_2) to which the charge is transferred befor recombination takes place. Now the molecular weight of this molecule is greater tha the atomic weight of the oxygen atoms which it is supposed are ionised by the inciden radiation ; it therefore follows that the scale height for the molecule is less than that fo the oxygen atoms and the gradient of the loss-coefficient is greater than that of the ionise gas. This state of affairs, envisaged by Bradbury, will, indeed, result in an increase i the electron concentration as we ascend above the level where the change-over occurs but this increase would continue, and there would be no resulting peak of concentratio in the F2 region unless some other cause produced one. What causes might there be

Of course, at great heights there is no atmosphere to ionise, and any vertica distribution of electron concentration must decrease to zero at the top. All reasonabl estimates of the air density show, however, that at the peak of electron concentratio only a small fraction (about 1 in 10^4) of the oxygen is ionised, and some other reaso must be sought for the existence of the peak.

It is now considered most probable that the F2 peak of electron concentration result from diffusion of the electrons through the neutral air under the action of gravity. Th electrons would, if left alone, diffuse faster than the positive ions, but the electrostati attraction between them causes both to diffuse with the same speed as determined by th ' ambi-polar ' diffusion coefficient, which is of the same order as the coefficient of the ion alone. This diffusion coefficient is inversely proportional to the gas density and is therefor more important at the greater heights. If electrons were suddenly inserted into the upp atmosphere with the same concentration at all heights, then in the absence of any proces which would remove them, they would be re-distributed by a process of diffusion, unti finally they took up the distribution appropriate to ' diffusion equilibrium,' in whic their concentration decreased exponentially upwards with the appropriate scale height They would, as it were, ' fall to the ground.' In the process of falling, however, th highest electrons, with the greatest diffusion coefficient, would fall most rapidly, so tha near the start of the diffusion process the high electrons would fall, the low ones would not and a peak of electron density would be formed as an intermediate stage in the process As time went on this peak would fall, until, in the end, it reached ground level.

In the ionosphere the circumstances are more complicated because there is always a process of electron-loss to be considered, and in addition, by day, there is a process of electron production. The argument above is sufficient to show, however, that diffusion can produce a peak of electron concentration, and it has been shown, particularly by Yonezawa, precisely how this can come about. The full calculation shows that the height of the peak is determined essentially by the ratio between the diffusion and the loss coefficients. To test these ideas it is reasonable to use the coefficients shown in Fig. 7 as a first approximation. To estimate the diffusion coefficient, it is necessary to postulate the air density, and this is one of the quantities which, up to the present, has not been well established. Recently, however, as a result of observations on artificial satellites, and with rockets, it has become reasonably certain that the air density at different heights lies somewhere in the region of the values shown in Fig. 8. If these values are used to compute the diffusion coefficient, and the loss-coefficients are taken from Fig. 7, then the theory of Yonezawa shows that the peak of electron concentration would be expected near a height of 300 km. This height is of the same order as the observed ones and it appears that we now have a first approximation to an explanation of the existence of the F2 layer.

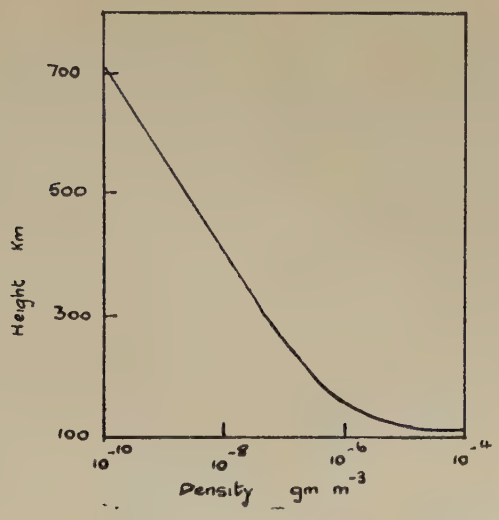

Figure 8. Air density at different heights estimated from rocket and satellite experiments.

We have now given, in outline, the present theory of the F2 layer and it is noticeable that it is quite different from the theory of the E and F1 layers. The F2 layer exists only because the loss coefficient decreases with height, as the result of a complicated charge-exchange process, and it has a peak only because diffusion becomes important at the greater heights.

It might be thought that once so much was clear it should be possible, by studying the behaviour of the layer at different parts of the earth, and at different times, to deduce the density of the ionisable gas, and the magnitude of the loss coefficient, at the different heights. Unfortunately, however, this is just where serious difficulties present themselves, for it is found that the behaviour of the layer is so unexpected that it is often described in terms of a series of ' anomalies ' which can be listed as follows :

(a) *An annual anomaly* which describes the fact that, over a large part of the earth in both hemispheres the peak electron density is greater than expected in December and January.

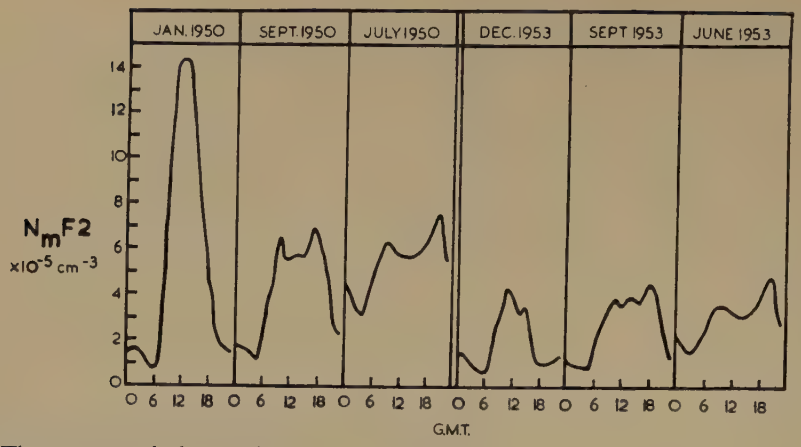

Figure 9. The average peak electron density of the F2 layer observed at Slough, on 10 International Geomagnetically Quiet days in each month. The ratio of the winter and summer values at midday is abnormal, but at midnight is normal.

(b) *A seasonal anomaly* which describes the fact that the peak electron density is greater in winter than in summer. In the Northern hemisphere the annual and the seasonal anomalies are additive and give rise to a striking result at Slough which is illustrated in Fig. 9. It is noticeable that in winter, in years of sunspot maximum, the peak electron density at midday is more than twice as great as in summer.

(c) *A diurnal anomaly* which describes the fact that the electron density at and near the peak reaches its maximum value at times far removed from midday, and often in the late evening. An example is shown in Fig. 10.

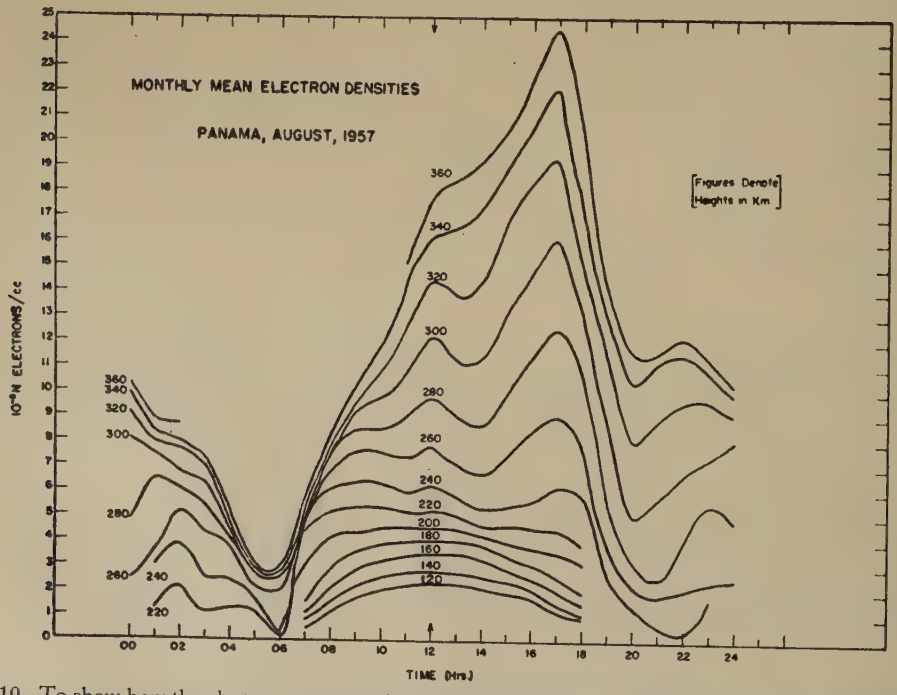

Figure 10. To show how the electron concentration varied with the time of day at a series of different heights above Panama in August 1957. The abnormal variation at the greater heights is noticeable as an increase between 14 and 18 hr.

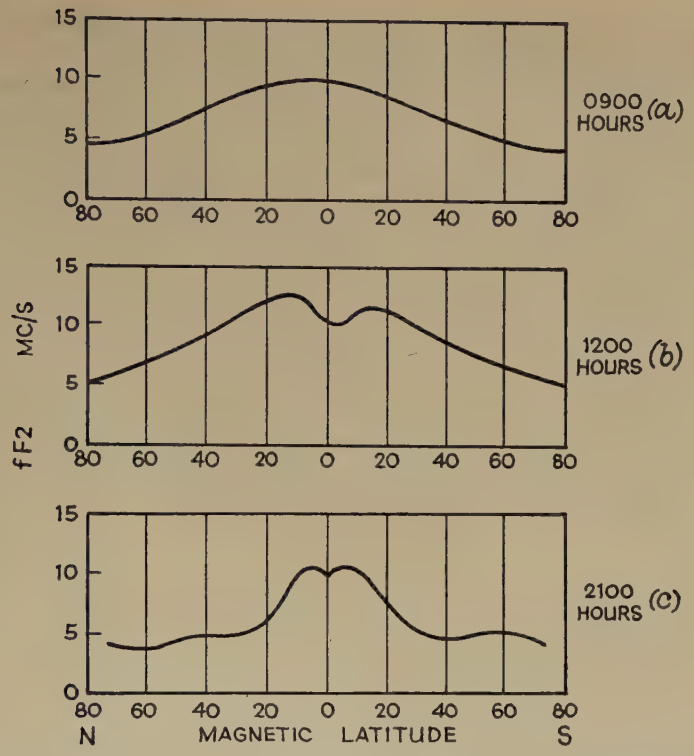

Figure 11. The critical frequency fF2, whose square is proportional to the peak electron density, plotted against magnetic latitude for three different hours at the equinox. A ' geographic anomaly ' is noticeable near the equator as a dip in the curve for 12 hr, and a peak in that for 21 hr.

(d) *A geographic anomaly* which describes the fact that, if the magnitude of the peak electron density at any one hour is plotted against latitude at the equinox, the resulting curve shows a dip or a rise near the equator according to the time of day selected. Results are more consistent if the magnetic dip is used instead of latitude in plotting the curve. Fig. 11 shows some results.

Although these anomalies may be caused, in part, by changes in the composition and distribution of the atmospheric gases from time to time and from place to place, it nevertheless seems fairly clear that they are largely the result of movements of the electrons in the F region. It is possible that the movements resulting from diffusion might be the cause of the geographic anomaly, since vertical diffusion is largely prevented near the equator, where the earth's magnetic field is horizontal. There is, however, the possibility of other important movements, related to the earth's magnetic field, of a kind which I shall now describe.

Movements of the electrons in the ionosphere can be the result either of movements of air molecules or of applied electric fields. When the air density is great it is comparatively easier to move the electrons by moving the air, but when it is small the electrons are moved comparatively more easily by the application of an electric field. In the denser E region, the air tides produced by the sun and the moon move the electrons, and this motion, in the presence of the earth's steady magnetic field, produces the currents responsible for changes in the geomagnetic field, and also produces a distribution of electric space-charge which in turn produces an electric field at the level of the F region. At that level, even if the air were moving, which is doubtful, the density is so low that the F region electrons are little affected by it and move more easily under the action of the imposed electric field and the permanent magnetic field. For the present discussion the movement in the vertical direction is the important one.

The theory of the ' atmospheric dynamo ' in the E region and the associated ' atmospheric motor ' in the F region has not yet been worked out in sufficient detail to determine properly how the vertical movements in the F region change with time or place over the earth, but sufficient has been done to show that the velocity could be large enough to alter the electron concentration as much as diffusion or electron-attachment. The movements are certainly of the right order to explain the anomalies, but no detailed account has yet been given to show why they occur as they do. It is, however, also clear that movements of the same order, varying as might be expected, regularly throughout the 24 hr, could seriously modify the ' normal ' behaviour of the layer.

The difficulties in solving the problem of the F layer now become apparent. Before its behaviour can be explained it is necessary to know the density of the ionisable gas; the density of the total atmosphere, which controls the diffusion; the density of the molecule which, through a process of charge-exchange, controls the loss coefficient; and the magnitude of the electron-velocities resulting from the applied electric field. At present it seems best to try to solve the problem by supposing that the total air density deduced from rockets and satellites is correct, so that the effect of diffusion is known, and using the ionospheric information to deduce the height distribution of the ionisable gas and the loss coefficient, and the magnitudes of the movements caused by electric fields.

In most of the work which has been done on this problem attention has been restricted to the variations of electron concentration at the peak of the layer, mainly because that is a quantity deducible directly from the easily measured critical frequency fF2. It has to be remembered, however, that the height of the peak is quite different at different times and places, and moreover, that the very existence of a peak results from a delicate balance between diffusion and electron-loss. The results are therefore difficult to interpret. Recently, several workers have deduced electron concentration profiles of the type shown in Fig. 2 for a series of places and times and from these it is now possible to construct and study curves which show the electron concentration (N) as a function of time (t) at fixed heights below the peak of the layer. The information derivable from these curves is complementary to that derivable from observations on the peak. An example was shown in Fig. 10. It is probable that, in the near future, the main features of that part of the F region below the peak will be explained, in outline, from a study of these data. It may take considerably longer to explain the anomalies in an acceptable way.

No theory of the F2 layer will, however, be complete until the electron distribution above the peak is known, and explained. Although the standard method of ionospheric sounding cannot be used to explore the upper part of the region, the method of measuring the Doppler shifted wave received from a rocket can be used, just as it is for the lower part of the ionosphere. Results for the upper part of the region are, however, few.

Other methods of exploring the upper part of the region include the measurement of the Doppler shift of waves received from an artificial satellite and a measurement of the Faraday rotation of the plane of polarisation of waves either emitted from an artificial satellite, or emitted from the earth and reflected from the moon. The Faraday rotation results from the bi-refringence of the ionosphere in the presence of the earth's magnetic field, and suitable measurements of it allow us to determine the total number of electrons in a unit column between the observer and the sender or the reflector.

Another valuable method of investigation makes use of the fact that electromagnetic waves of audio-frequency can travel through the ionosphere guided along the magnetic lines of force, and with a velocity which depends on the frequency of the wave and the concentration of the ionospheric electrons. When, therefore, powerful waves of these frequencies are radiated by a lightning flash they enter the ionosphere and are guided back to a point on the earth at the other end of the appropriate line of magnetic force. At the far end, however, the higher frequencies arrive first, so that if the wave is received on an aerial, amplified, and applied to a loudspeaker, a whistle of descending pitch is heard. From the relative delay of the different frequencies in this whistle, and a knowledge

of the shape of the line of force, it is possible to make an estimate of the electron concentration at the greatest radial distance reached by the line of force. By choosing observing points at different latitudes, corresponding to lines of force which have reached to different radial distances, estimates have been made of the electron concentrations at considerable distances from the earth.

The results of experiments made to measure the electron concentration in the outer parts of the F region are collected in Fig. 12. Although these are tentative and somewhat inaccurate, they show, fairly conclusively, that the electron concentration falls off with increasing radial distance less rapidly than might have been expected. Near the peak the gradient is about half that expected, and far out it is smaller still.

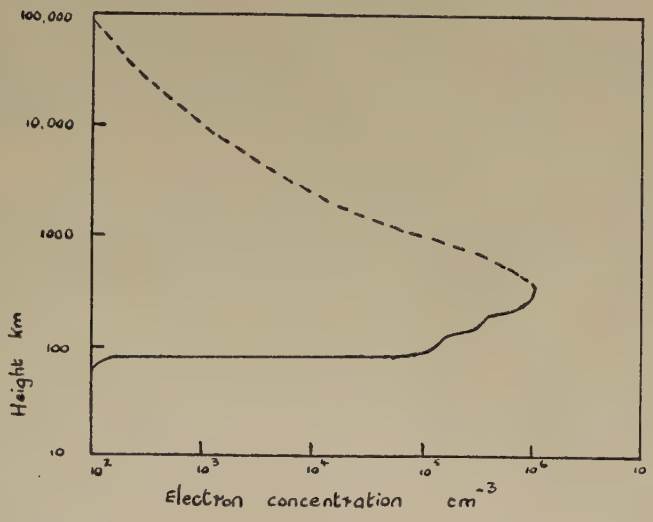

Figure 12. A rough estimate of the electron distribution on the upper side of the ionosphere, based on experiments with rockets, satellites, and ' whistlers.'

Chapman has recently suggested that the outermost parts of the ionosphere merge continuously into the outmost parts of the solar corona and that the electron concentration in the corona, at the distance of the earth, may be as great as 1,000 cm^{-3} and the temperature as high as 10,000°K. If this suggestion were correct the upper part of the curve in Fig. 12 might be determined more by the corona than by the earth's atmosphere. There would also be a flow of heat downwards, through the atmosphere, from the hot corona above, with the result that in the upper part of the ionosphere there might be a continuous gradient of temperature. A gradient of this kind is, indeed, required to explain the comparatively great air densities revealed by artificial satellites at the greater heights.

Ten years ago it was widely thought that the F2 layer of the ionosphere was formed, like the E and F1 layers, by a simple process of the type described by Chapman long ago. Five years ago it was beginning to be realized that, quite apart from the problem of explaining the ' anomalies,' an explanation of this type was impossible, but it seemed that there were so many ' unknowns ' in the problem that a unique explanation could not be given. Recently, however, the determinations of atmospheric and electronic densities by rockets and artificial satellites, and the calculations of electron density distributions from the h′ (f) curves, have provided just enough extra information to encourage the belief that a first-order solution of the problem may not be long delayed. Once the first-order solution is available the way will be clear for a consideration of the anomalies in the F2 region, and of the fundamental changes noticeable in it during ionospheric storms.

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World temperatures during the Pleistocene

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(Manuscript received 16 January 1959)

Summary

In previous papers it has been shown that the glacial and pluvial periods of the Pleistocene can b explained qualitatively on the assumption that during this period there were three oscillations of sola radiation. Accepting this assumption the present day zonal temperatures – mean temperatures along circle of latitude – are considered and the physics of the relationship between changes in solar radiatio and the consequent changes in the zonal temperatures are discussed. By a graphical method of trial an error a family of curves relating the zonal temperatures of all latitudes with the oscillations of the solar radiatio is prepared and discussed. These results are tabulated and used to plot curves showing the variations of th zonal temperatures of five latitudes between the Equator and the North Pole during a complete cycle of sola radiation and applied to the glacial and interglacial periods.

The difference between the actual temperatures of any place and the zonal temperature of the latitud on which it is situated – the so-called temperature anomaly – is then discussed and two diagrams ar prepared showing the local temperatures at places in the British Isles and North America during the las cycle of solar radiation embracing the period from the beginning of the Mindel-Riss interglacial to the presen day. The geological evidence for the temperatures thus derived is briefly discussed.

1. Introduction

In two previous papers published by the Royal Meteorological Society in 1934 and 1957 I described a theory of climate change during the Pleistocene based on the assumption of periodic variations of solar radiation. It was postulated that an increase of solar radiation (solar constant) above its present-day value would raise the temperature of the Earth as a whole; it would also lead to an increase in the temperature difference between the Equator and the Poles which would be accompanied by an increase in the activity of the general circulation of the atmosphere. The increases in the temperature and the circulation would lead to increased evaporation and consequently to increased precipitation; which in high latitudes would be mainly in the form of snowfall. This would lead to the formation of polar ice-caps which would extend outwards until they melted in the higher temperatures of the lower latitudes. This would be the maximum of a glacial period.

As the solar radiation continued to increase and the temperature outside the ice-cap to rise, the margin of the ice-cap would retreat by melting, in spite of continued increased precipitation. By the time the radiation reached the maximum of the cycle the polar ice-cap would have disappeared and the precipitation would be much heavier than at present in all latitudes; the earth would then be experiencing a pluvial period. When the crest of the cycle had passed the radiation would decrease and the changes in the temperature and precipitation would be repeated in the reverse order passing through a glacial period* on the way to the minimum radiation.

The results described in the previous papers are summarized in Fig. 1 which is based on Fig. 3 of the 1957 paper. The diagram extends over the million years from the beginning of the Pleistocene period to the present day as shown by the scales in the divisions numbered 1 and 9 in the column on the left of the diagram. The curve in divisions 4-7 represents the postulated variation in the intensity of the solar radiation (solar constant). It is a simple periodic variation with approximately two and a half oscillations during the whole of the Pleistocene Period. As the solar radiation varies so the temperature of the Earth rises and falls with it, and the other climatological factors vary in consequence.

* Geologists use the expressions ' glacial ' and ' pluvial ' with their complementary ' interglacial ' and ' interpluvial ' periods; but no quantitative definitions are given. In my 1957 paper (p. 467) I defined a glacial period by the phrase : ' *a glacial period exists whenever the margin of the north polar ice-cap extends on the average south of latitude 75°N,*' which is its approximate position at the present day. This definition is used throughout this paper.

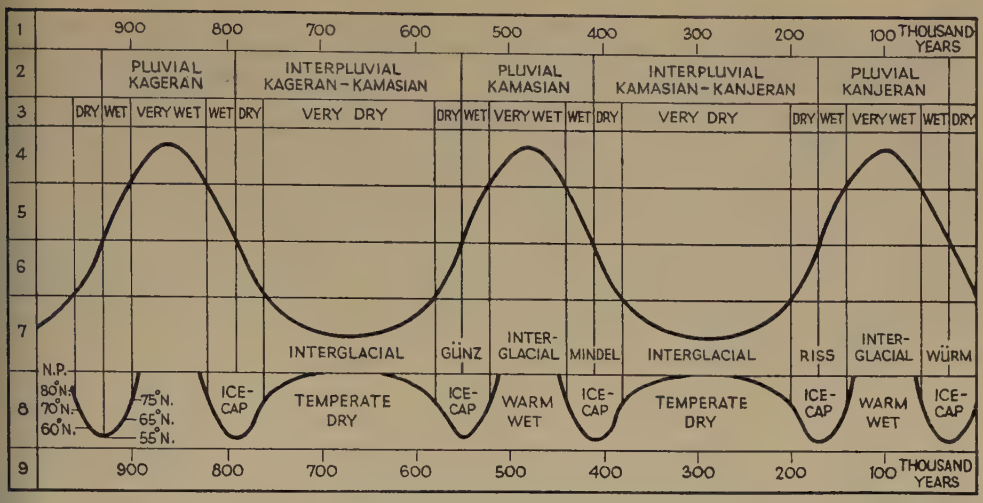


Figure 1. The radiation curve.

Thus the curve of radiation in Fig. 1 is also a curve of temperature, but it is only a ualitative one, for in the previous papers it was not found possible to obtain a numerical elationship between the radiation and the temperature. It is the object of this paper to nd this relationship which will obviously be different in different parts of the world.

In Fig. 1 the vertical lines below the radiation curve mark the glacial periods, six of vhich are shown during the Pleistocene, but only four are named, these are the glacial eriods in the Alps named by Penck and Brückner in 1909, Günz, Mindel, Riss and Würm; he last three are generally accepted by geologists as occurring in the last half of the 'leistocene. There is geological evidence in Europe for at least one glacial period before he Günz (the Donau Gravels) and probably for others; but as the period and extent of hese early glaciations have not been determined by geologists names are not assigned to he two early glacials in Fig. 1. The pluvial and interpluvial periods are indicated by ertical lines above the radiation curve and the names of three pluvial periods are entered n division 2. The curve in division 8 shows the latitude of the margin of the polar ice-cap s it advances and retreats.

2. Present-day temperatures

The actual mean temperature of a locality is very largely affected by height above ea-level and by its geographical surroundings, especially whether these are continental r oceanic; but when the mean monthly or annual sea-level temperatures over each circle f latitude are computed we obtain a series of ' zonal temperatures ' which are remarkably egular from latitude to latitude. The present-day zonal temperatures for January and uly, at each five degrees of latitude in the Northern Hemisphere, due to Meinardus Hann-Suring 1926, pp. 138-140) are given in Table 1, column II, p. 338.

The explanation of the present-day distribution of terrestrial temperature may be lescribed in broad outline in a few words. Owing to the shape of the Earth, more solar adiation is received per unit area in equatorial than in polar regions; this produces a emperature gradient between the equator and the poles which in turn produces differences n barometric pressure. The resulting air-movement known as ' the general circulation f the atmosphere ' leads to a mixing of the air from high and low latitudes until a emperature distribution is reached in which the total energy received by the Earth as olar radiation is returned to space as terrestrial radiation. The present-day zonal emperatures from Table 1, column II are shown by the thick curves in Fig. 6 (a), p. 347 nd represent this balance at the present day. The most important characteristic of these urves, from our point of view, is the regularity or smoothness of the curves from latitude

to latitude all the way from the Equator to the Pole. Land, sea, mountain ranges an oceanic currents have a large influence on the temperature of the air and they vary large from latitude to latitude; yet there is no indication of them in the zonal temperature it is therefore legitimate to assume that surface features are not important in determinin the radiation balance of the atmosphere. There is, however, one surface feature whic plays a powerful part in climate and that is a surface of snow and ice under certa circumstances. This will receive separate consideration in a later section.

With constant solar radiation the general circulation of the atmosphere would rema the same in corresponding seasons from year to year, except for minor irregulariti probably due to aerodynamic instability which do not concern us here. Our proble now is to determine how this steady state of the atmosphere reacts to variations in th intensity of the solar radiation.

3. Temperatures and changes in solar radiation

Geological evidence is clear that the Earth has recently passed out of a glacial into a interglacial period so that the present day is at the division between the two phases. Th existing small ice-cap at lat. 75°N will decrease and disappear during a cold-dry interglaci similar to the Mindel-Riss interglacial. In view of the remark in a previous paragrap that snow and ice under certain conditions play an important part in climate this chapt will be divided into two sections : the first dealing with the interglacial periods when the is so little snow or ice that its effect may be neglected and the second with the glaci periods during which the climate in high latitudes is dominated by the polar ice-cap

(*a*) *Interglacial periods*

The primary effect of an increase in solar radiation is to raise the temperature of th Earth as a whole, and when the steady state is re-established the increase would be great in equatorial than in polar regions. Expressed in terms of zonal temperature, the zon temperatures of every latitude would be raised and the temperature difference betwee successive latitudes would be increased. We have no means of calculating what wou be the increase in the temperature gradient. To make our problem at all tractable th theoretical change in the gradient will be neglected. Thus during interglacial perio when ice has little or no effect on the temperature a given change in solar radiation w be assumed to produce the same rise or fall in the zonal temperature in all latitudes. Th includes the Equator; now at the Equator there is no annual variation of the temperatur therefore any change in equatorial temperatures due to changes in solar radiation mu be the same in all months. This involves that during interglacial periods the changes the zonal temperatures due to changes in solar radiation must also be the same in a months (summer or winter) in all latitudes.

Thus if E is the equatorial temperature at the present time and E' at any other specifie interglacial phase, and Z and Z' are the corresponding zonal temperatures of any give latitude, so that $Z' - Z$ and $E' - E$ are the changes in the zonal temperature at the chose latitude and at the Equator, then as $Z' - Z = E' - E$ we have $Z' = Z + (E' - E$ or in a more convenient form $Z' = Z + \Delta E$, in which ΔE is the change at the Equato As we shall be using graphical methods in what follows it will be advantageous to s what this means when applied to our diagrams.

In Fig. 6 (a) (p. 347) the thick curves represent the January and July zonal tempe atures for all latitudes at the present day. The two curves commence together at th Equator at 26°C. Now let the radiation be reduced until the temperature at the Equat falls to 24°C. The temperature at every other latitude falls by two degrees also and th result is a pair of curves two degrees below the original pair at every point. Thu increasing or decreasing the radiation during an interglacial period results in a new pa of curves parallel to the original curves; but raised or lowered by the same amount as the Equator.

(b) *Glacial periods*

Considering periods of the order of a month or so the mean temperature of the ground is near to the mean temperature of the air in contact with it. This condition applies also to a surface of ice so long as the temperature of the air is below the freezing point. When the temperature of the air is above 0°C the surface temperature of the ice rises to the melting point and cannot rise further. The air, however, continues to lose heat to the ice by radiation and contact and to the upper air by radiation, so that the temperature of the air falls, and if it continues sufficiently long in contact with the ice its temperature also will fall to 0°C. This property of ice plays an important role in the control of the air temperature during glacial periods. As the effect is different in summer and winter it will be necessary to consider the winter (January) and the summer (July) seasons separately. We shall also confine the discussion to the maximum glacial phase, when the ice-cap is at its maximum extent.

(c) *Maximum glacial phase : January*

According to our definition (p. 332) the margin of the ice-cap at the beginning of a glacial period is on the average at lat. 75°N. It was shown in my 1957 paper that during each glacial period the ice-cap extends to lat. 55°N then retreats and reaches lat. 75°N again at the end of the period. Outside the polar ice-cap during this period the snow disappears everywhere at sea-level during the summer months, but during the winter months for several degrees of latitude from the ice-cap there would be more or less snowfall and the snow would lie unmelted for longer or shorter periods. During these periods in the winter when the snow lies outside the ice-cap the air temperature may be either above or below the freezing point. As we have already seen when the air temperature is below 0°C the ice surface has no effect on the air temperature. On the other hand, when the air temperature is above 0°C with snow lying the snow melts and disappears. The surface then regains its normal state and takes up the temperature of the air. The result is that outside the limits of the ice-cap temporary periods of snow-cover during the winter have little or no effect on the zonal temperatures. This conclusion is illustrated by the following example.

Every year the whole of the Northern Hemisphere north of lat. 50°N, except the northern parts of the Atlantic Ocean, has a covering of snow throughout the whole of January so that there is a rapid change in the nature of the surface near lat. 50°N from a snow-free to a snow-covered surface; yet the curve for the January zonal temperature at the present day in Fig. 6 (a) shows no discontinuity or change of slope (temperature gradient) where the snow-free changes to a snow-covered surface in the neighbourhood of lat. 50°N. It will be noticed that the January zonal temperature in this area is well below the freezing point.

The temperature over all parts of the ice-cap is below freezing throughout January (otherwise, as January is the coldest month, there would be no ice-cap at all) therefore in January the ice-cap itself does not affect the surface temperature and what has been said above about the conditions outside the ice-cap will apply over the whole of the ice-cap also during the month of January and we get our final result :

In January, neither over nor outside the polar ice-cap does a snow surface have any influence on the radiation balance; therefore in January the zonal temperatures over the whole of the Northern Hemisphere change with the radiation so that the equation $Z' = Z + \Delta E$; which we have already found for all months and all latitudes during the interglacial periods (see p. 334), applies also during January to the glacial periods. This does not apply to the glacial periods in the summer; therefore it is necessary now to consider the zonal temperatures during the glacial periods in the month of July.

(d) *Maximum glacial phase : July*

In order to discuss the July zonal temperatures at the maximum of the glacial phase it will be convenient to divide the Northern Hemisphere into three zones : (1) an equatorial

zone from the Equator to lat. 35°N, (2) an ice-cap zone from lat. 55°N to the North Pole and (3) the intermediate zone between (1) and (2). (See vertical divisions in Fig. 6).

(1) The equatorial zone, Equator to lat. 35°N : It will be seen from the upper thick curve in Fig. 6 (a), curve C, that the July zonal temperature of lat. 35°N at the present day, is 26°C, the same as at the Equator, and that at lat. 20°N, the temperature is two degrees higher. This is because in July the declination of the sun is approximately 20°N so that at lat. 20°N the mid-day sun is in the zenith and the maximum amount of solar radiation is then received along this latitude. For our purpose it is important to notice that in the equatorial zone it is the position of the sun which determines the temperature in July. It is reasonable, therefore, to assume that the distribution will be similar for all values of the solar constant. Hence in this zone at all times the temperature of the zone will rise and fall with the radiation, so that in the equatorial zone the temperature of any latitude, Z', will be given by $Z' = Z + \Delta E$, the expression we have found for all latitudes during interglacial periods. (§3 (a)). It applies also to the equatorial zone during glacial periods because the ice-cap is too far away to influence the equatorial temperatures.

(2) The ' ice-cap zone,' lat. 55°N to the North Pole : At the maximum of a glacial period the ice-cap, after extending from the North Pole to lat. 55°N, is temporarily in equilibrium before retreating. It is not difficult to make a reasonable estimate of the temperature at the margin of the ice-cap in this phase. The physical conditions at the edge of every polar ice-cap are practically the same during the summer months : a region having a snow surface on one side of the margin with temperatures decreasing towards the pole; and a region free from snow on the other side with temperatures increasing as one recedes from the ice-edge. The position of the margin when in equilibrium is determined chiefly by the air temperature; it does not vary appreciably with the seasons nor from year to year, we may therefore assume that the thermal conditions during July are similar at the edge of all polar ice-caps. It seems reasonable to assume that if we know the average July temperature at the margin of any one polar ice-cap we may use that as representative of all other ice-caps.

Unfortunately the two existing polar ice-caps are not satisfactory from this point of view. The margin of the present Antarctic ice-cap is indeterminate : the ice extends from an extensive plateau to the coast and covers the sea in great ice shelves, the margins of which are constantly breaking off and floating away; nowhere in the Antarctic is the margin of the ice-cap mainly determined by the local air temperature, as it would be on a continent distant from the ocean.

The conditions are not much better in the Arctic for our purpose. The present north polar ice-cap is very small and far from circular. We have, however, decided to consider it is equivalent to an ice-cap centred on the pole and extending to lat. 75°N. (Simpson 1957, p. 425). From Table 1, column II (p. 338) we find that at the present day the July zonal temperature at lat. 75°N is 4°C and that 0°C is reached in lat. 85°N, that is 10° of latitude (1,000 km) within the margin. These figures are only approximate and not entirely appropriate; but for want of other data I adopt them for all the Pleistocene north polar ice-caps irrespective of size; in any case no other reasonable figures would seriously affect the argument. Thus the zonal temperatures for July over the north polar ice-cap, irrespective of size and phase are taken to be 4°C at the margin, 0°C ten degrees of latitude nearer the pole and then continuing to decrease slowly to the pole.

(3) The ' intermediate zone,' lat. 35°N to lat. 55°N : In this zone there is no permanent ice, so ice cannot have any immediate effect on the air temperature; but at the maximum of the glacial phase the margin of the ice-cap is at lat. 55°N, bringing with it a July temperature of 4°C; this is many degrees lower than it would be in the absence of the ice-cap. This depression of the temperature at the margin of the ice-cap, as we shall see in § 5 III, lowers the July zonal temperatures in the ' intermediate zone ' but its influence does not extend so far as the equatorial zone.

We have now gathered a considerable amount of information about the temperatures in the Northern Hemisphere during the glacial and interglacial periods which is co-ordinated and summarized in the two following lemmata.

(*e*) *Lemma I. All interglacial phases*

The zonal temperature, Z', of any latitude at a specified interglacial phase is equal to the present day zonal temperature, Z, plus ΔE, in which ΔE is the difference between the equatorial temperature at the present day and at the specified phase, i.e., $Z' = Z + \Delta E$ (§ 3 (a)).

(*f*) *Lemma II. Glacial phases*

(*a*) *All glacial phases :* The zonal temperatures for January at all latitudes are given by $Z' = Z + \Delta E$ (§ 3 (c)).

(*b*) *In the ' equatorial zone '* (Equator to lat. 35°N) the rule $Z' = Z + \Delta E$ applies to all months in the glacial as well as in the interglacial periods (§ 3 (d.2)).

(*c*) *In the ' intermediate zone '* (lat. 35°N to lat. 55°N) there is some slight lowering of the present-day July zonal temperature at the maximum of each glacial phase. (§ 3 (d.4)).

(*d*) *In the ' ice-cap zone '* (lat. 55°N to the North Pole) : At the maximum glacial phase (equatorial temperature 32°C) the July temperature is 4°C at the margin of the ice-cap (lat. 55°N) and 0°C ten degrees of latitude nearer the pole (lat. 65°N). (§ 3 (d.3)).

4. Table of zonal temperatures for all latitudes during a complete cycle of solar radiation

Our next step is to combine all this information into a consistent description of the climate during the Pleistocene Period giving actual values of the zonal temperature of all latitudes in all phases of the glacial and interglacial periods.

It is quite impossible to do this by strict mathematical methods; but a great deal can be done by graphical methods with the aid of ' trial and error.' After several trials it was found possible to construct four diagrams containing the required information. In the Appendix the method used is described and the diagrams are reproduced. Here it is only necessary to refer to Table 1 which has been prepared from the diagrams to give the results in a more convenient form for further discussion.

Table 1 (p. 338) contains the January and July zonal temperatures for every fifth degree of latitude for five principal glacial and interglacial phases of a complete cycle of solar radiation. The zonal temperatures are contained in five columns each giving the January and July zonal temperatures with their arithmetical means, the latter being approximately the mean annual temperature. Contents of each column will be found at the head of the column.

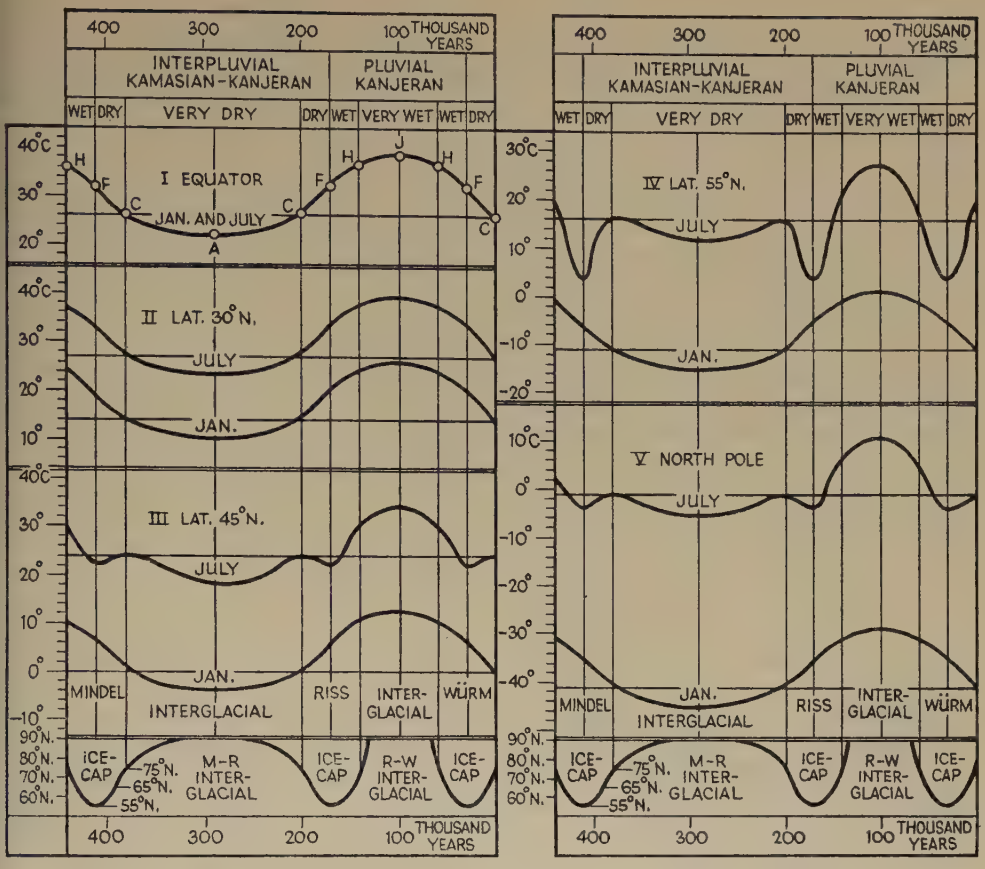
5. Temperature variations of five latitudes during a complete cycle of solar radiation

We now have the data to study the changes in the zonal temperature of any latitude as the solar radiation goes through a complete cycle and to compare these changes with the simultaneous changes in the various climatic zones from the Equator to the North Pole.

This can best be done by plotting curves of zonal temperature for January and July (as representing the winter and summer seasons) against time for a number of typical latitudes in such a way that they can be compared, and then to discuss the curves with reference to the glacial and interglacial phases. Such a set of curves for five latitudes are shown in Figs. 2 (a) and (b) (p. 339).

TABLE 1. ZONAL TEMPERATURES AND CLIMATIC PHASES

Column	I			II			III			IV			V		
Phase	(A) Min. Radiation			(C) Present day			(F) Max. glacial			(H) Ice-cap at lat. 75°N			(J) Max. Radiation		
Radiation increasing	Mid M-R			End M and W			Mid Riss			End Riss			Mid R-W		
Radiation decreasing	Mid M-R			Beginning R			Mid M and W			Beginning M and W			Mid R-W		
Derived from	Col. II − 4°C			Meinardus			Fig 6 (b)			Col. II + 10°			Col. II + 12°C		
	Jan. °C	Mean °C	July °C	Jan. °C	Mean °C	July °C	Jan. °C	Mean °C	July °C	Jan. °C	Mean °C	July °C	Jan. °C	Mean °C	July °C
Equator	22	22	22	26	26	26	32	32	32	36	36	36	38	38	38
5°N	22	22	22	26	26	26	32	32	32	36	36	36	38	38	38
10°N	22	22	23	26	26	27	32	32	33	36	36	37	38	38	39
15°N	20	22	24	24	26	28	30	32	34	30	34	38	36	38	40
20°N	18	21	24	22	25	28	28	31	34	32	35	38	34	37	40
25°N	15	20	24	19	24	28	25	30	34	29	34	38	31	36	40
30°N	10	17	23	14	20	27	20	26	33	24	30	37	26	33	39
35°N	6	14	22	10	18	26	16	24	32	20	28	36	22	30	38
40°N	1	10	20	5	14	24	11	20	28	15	34	33	17	26	36
45°N	− 4	7	18	0	11	22	6	13	20	10	20	30	12	23	34
50°N	− 10	2	15	− 6	6	19	0	4	9	4	14	24	6	18	31
55°N	− 15	− 2	12	− 11	3	16	− 5	− 1	4	− 1	9	20	1	14	28
60°N	− 20	− 6	9	− 16	− 2	14	− 10	− 11	2	− 6	5	17	− 4	12	25
65°N	− 25	− 10	6	− 21	− 6	10	− 15	− 8	0	− 11	2	16	− 9	6	22
70°N	− 30	− 14	3	− 26	− 10	7	− 20	− 10	− 1	− 16	− 2	12	− 14	2	19
75°N	− 34	− 17	0	− 30	− 13	4	− 25	− 14	− 2	− 20	− 5	10	− 18	− 1	16
80°N	− 37	− 20	− 2	− 33	− 16	2	− 29	− 16	− 3	− 23	− 7	8	− 22	− 4	14
85°N	− 41	− 22	− 4	− 37	− 18	0	− 32	− 18	− 4	− 27	− 10	7	− 25	− 6	12
North Pole	− 45	− 25	− 5	− 41	− 21	− 1	− 35	− 19	− 4	− 31	− 13	6	− 29	− 9	11



(a) Equator to lat. 45°N. (b) Lat. 55°N to North Pole.

Figure 2. Zonal temperatures and climatic phase for five latitudes.

Fig. 2 is divided into two parts : (a) contains the curves for I the equator, II lat. 30°N and III lat. 45°N; and (b) contains those for IV lat. 55°N and V the North Pole. At the head of each diagram the limits of pluvial and interpluvial phases are given and at the foot those of the glacial and interglacial phases with a curve of the advance and retreat of the polar ice-cap. Scales of years before the present are given in the top and bottom lines and scales of temperature in degrees centigrade are at the left of the curves for each of the chosen latitudes.

The period discussed is from the beginning of the Mindel glacial to the present day at the end of the Würm glacial. This comprises a little more than one complete cycle of solar radiation. The data for each curve are taken from Table 1 (p. 338).

Each curve ends on the right of the diagram at the present-day zonal temperature for the latitude represented. From this point a horizontal line is drawn across the diagram to give a base-line from which the departures of the zonal temperatures above or below the present-day temperatures can be seen at a glance.

I. The Equator. The curve for the Equator is a copy of Fig. 5 (p. 346) the construction of which is described in the Appendix. As there is no annual variation of temperature at the Equator one curve is sufficient to represent the temperature of both January and July.

The curve in the diagram shows that at the end of the Mindel glacial, which is the beginning of the Mindel-Riss interglacial also, the equatorial temperature is 26°C, and is slowly falling. In the middle of this interglacial the radiation cycle reaches its

minimum and the equatorial temperature was then 4° below its present-day value. Th radiation then commenced to increase and at the end of the interglacial, the temperatur was back at 26°C as at the beginning. Thus at the Equator the lowest temperature of th whole radiation cycle was contemporary with the middle of the Mindel-Riss interglaci in high latitudes and with the minimum precipitation of the Kamasian-Kanjera interpluvial.

After the interglacial the temperature continued to rise until the radiation maximur was reached in the middle of the Riss-Würm interglacial with an equatorial temperatur of 38°C, 12° higher than at the present day. It was also coincident with the middle of th Kanjeran pluvial period.

From this maximum the radiation and the temperature decreased and present-da temperatures were reached at the end of the Würm glacial.

Thus during one cycle of radiation the equatorial temperature goes through on oscillation having minimum and maximum temperatures coinciding with minimum an maximum precipitation in the interpluvial and pluvial periods respectively, without an indication on the temperature curve of the Riss and Würm glacial periods.

II. Latitude 30°N. There is an annual variation of temperature at lat. 30°N, so tw curves are given for this latitude, one for January with a base-line at 14°C and one fo July with a base-line at 27°C. The curves are both parallel to the curve for the Equato and therefore the zonal temperatures for all latitudes from the Equator to lat. 30°N ris and fall by the same amount as the temperature at the Equator. The area of the Earth' surface from the Equator to lat. 30°N is half the total surface area of the Northern Hemisphere; hence we have the important result that the zonal temperatures over th equatorial half of the Northern Hemisphere rise from a minimum during the Mindel Riss interglacial to a maximum during the Riss-Würm interglacial; but over the whol of this great area the temperature shows no sign of the glacial periods which are such a marked feature of the climate in the other half of the hemisphere; on the contrary, a the time when the ice-cap is at its greatest extent the whole equatorial zone from th Equator to lat. 35°N is 6° warmer than at the present day.

III. Latitude 45°N. At this latitude we see the first signs of the effect of the polar ice the curve for the July temperatures shows a depression during each glacial period. At thi time the ice-cap had reached its maximum extent with its margin at lat. 55°N where, a we see from Table 1, column III, the ice had lowered the July temperature to 4°C, which is 12° below its present day value and 18° below what it would have been at that time i there had been no ice-cap. It is this lowering of the July temperature by the proximity of the ice-cap which is the cause of the drop in temperature at the maximum phase of the glacial periods in the ' intermediate zone.' (see § 3 (d.3) p. 335).

The January curve does not show a similar dip, but rises and falls with the temperature curve for the equator. It will save repetition to remind the reader here that according to lemma II (a) (p. 337) the curve for the January zonal temperature *for all latitudes* is parallel to the curve for the Equator and therefore retains its simple periodic shape with one maximum and one minimum each coinciding with an interglacial phase; also the January temperature in all latitudes at the maximum of each glacial period is 6° higher than at the present time.

IV. Latitude 55°N. At this latitude we first meet the polar ice-cap itself which reaches lat. 55°N at the maximum of each glacial phase.

The January zonal temperature as usual follows the equatorial temperature and oscillates between — 15°C in the Mindel-Riss interglacial and + 1°C in the Riss-Würm interglacial. It is below its present-day temperature only during the Mindel-Riss interglacial and then by 4° at the most; during the rest of the radiation cycle it is above its present-day value – by 6° at the maximum glacial phase and by 12° in the middle of the Riss-Würm interglacial.

On the other hand, the July zonal temperature which at the beginning of the Riss glacial phase is 16°C, as at the present day, is brought down 12° to 4°C when the margin of the ice-cap arrives in the middle of the glacial phase. This is the direct effect of the

local ice on the July temperature. As the solar radiation increases the mean temperature rises and the ice-cap retreats through melting. When the ice-cap has returned to 75°N the glacial period is considered to end, and the further retreat is part of the Riss-Würm interglacial period.

After passing the maximum radiation the July temperature falls. An ice-cap forms at the pole and extends to lat. 75°N when the Würm glacial commences; the ice-cap continues to grow, reaches lat. 55°N with a temperature of 4°C and then retreats and is back again at lat. 75°N at the present day.

V. The North Pole. The curve for the July zonal temperature at the North Pole shows a considerably smaller depression of the temperature during the glacial period than occurred at lat. 55°N when the margin of the ice-cap reached that latitude. At the present day the July temperature at the North Pole is practically at 0°C and during the whole radiation cycle it does not vary more than 3-4° from this value except during the Riss-Würm interglacial when the radiation is at its maximum, and the July temperature rises to 11°C with a mean annual temperature of — 9°C. Whether this high summer temperature resulted in the disappearance of the whole remnant of the ice-cap we have no evidence; but the winter temperature at the pole was then — 29°C and the precipitation at its maximum. The accumulation of snow during the winter must have been very great so that it is possible that the ice-cap did not entirely disappear during the summer every year.

6. Local temperatures

(*a*) *Temperature anomalies*

So far we have dealt only with zonal temperatures and we have devised methods for obtaining zonal temperatures for all latitudes at all phases of the glacial and interglacial periods. Zonal temperatures however are not sufficient for specifying the temperatures in individual localities, for the actual temperatures in many parts of the world depart largely from the mean temperature of the latitude, especially when we consider seasonal values. For example, in NE. Asia in lat. 65°N, long. 130°E the actual average January temperature is more than 20°C below the January zonal temperature; while off the coast of Norway in the same latitude and month the local temperature is more than 20° above the zonal temperatures : that is a difference in temperature of over 40°C between two localities on the same latitude and in the same month. Similar differences, but not so large, occur also in the summer : the July temperature in North Africa (the Sahara) is more than 8° above, and over the NE. Pacific Ocean more than 8° below the zonal temperature. The difference between the actual temperatures of a locality and the mean temperature of the latitude is called the temperature anomaly. Hann-Suring (1920, Figs. 18 and 19) gives world charts of temperature anomalies for January and July in which the distribution of the anomalies is shown by lines called isanomalies. From these charts it is easy to see the causes of the anomalies in the examples given above : the negative anomaly over NE. Asia in January is due to ' continentality,' that is winter temperatures over land, especially in high latitudes, are lower than over oceans on the same latitude; the positive anomaly off the coast of Norway is due to the Gulf Stream in the Atlantic, and the one over the Sahara is due to continentality and the sand surface of the desert. These are all important geographical features capable of affecting the local climate.

(*b*) *Temperature anomalies and solar radiation*

From what has already been said it will be clear that every temperature anomaly is the effect of one or more geographical feature acting on the ideal temperature distribution which would be set up by solar radiation on the Earth if its surface were uniform. The existing zonal temperatures approach that ideal distribution with the present solar constant. The geographical features which cause temperature anomalies, such as the distribution of land and sea, mountain ranges and oceanic currents are themselves permanent features

of the physical surface of the Earth and would produce similar local variations of temperature cloud, wind direction and rainfall whatever might be the temperature of the world as whole within the limits of the variations in solar radiation with which we are dealing We may therefore assume that the value of an anomaly will remain unchanged as the zonal temperature rises and falls in consequence of changes in the solar constant unless the geographical features which cause the anomaly undergo a material change; as for example when an ice-cap overrides a region which at other times is ice-free.

As pointed out above, the existing zonal temperatures represent the thermal balance with the present solar constant; the zonal temperatures at other climatic phases set out in Table 1 correspond to the different solar constants at those phases; therefore it is only necessary to add the present-day temperature anomalies to the zonal temperatures in Table 1 to have the local temperatures at the latitudes and phases included in the tables and at all other latitudes and phases by interpolation.

In the few cases in which present-day anomalies are subject to change, usually as the result of the advance and retreat of the polar ice-caps during the glacial periods, it is often possible to obtain the local temperature on physical and meteorological considerations. Such a case for England is illustrated in § 7.

(*c*) *Local temperature and sea-level*

So far in this paper all the temperatures discussed have been temperatures at sea-level. Local temperatures are, however, as their name implies, the actual temperatures at the place of observation, and therefore are not comparable with zonal temperatures unless the height above sea-level is taken into account. The simplest method to do this is to treat the temperature difference due to the height above sea-level as an anomaly and to add it to any other anomaly. This is the method used in preparing isothermal charts, the local temperatures being corrected by adding the sea-level anomaly before being used in the chart.

The rate at which temperature decreases with height depends on a number of factors but for climatological work it is sufficiently accurate to deduct 0·5°C for each 100 metres of height (1·1°F for each 400 ft). This factor is used in the preparation of practically all text-books of climatology.

7. Local temperature in the British Isles

Theoretically we are now in a position to derive the local temperature in the Northern Hemisphere at any climatic phase during a complete cycle of radiation. As illustrations of the method the local temperatures of two localities will be discussed in detail. For one of these I have chosen the British Isles (lat. 55°N, long. 0°) : at the present day this region has a very large January temperature anomaly due to the Gulf Stream which is swept away when the ice-cap arrives. Also a great deal of geological and botanical research has been done in recent years on Pleistocene glacial and interglacial deposits in England especially in East Anglia, which is useful for testing theoretical results.

The first step is to prepare a diagram similar to those already used in Fig. 2, but with local temperatures replacing zonal temperatures. Fig. 3 has been prepared for this purpose. The lay-out is similar to that of Fig. 2 and needs no further explanation.

(*a*) *Local temperatures during the interglacial periods*

We have already seen that provided there has been no appreciable change in the geographical surroundings the local temperature at any place during all interglacial periods was the zonal temperature at that time and place plus the present-day anomaly. We are dealing now with the last two interglacial periods and there can be no doubt that there has been no major change in the geographical conditions of the North Atlantic in that

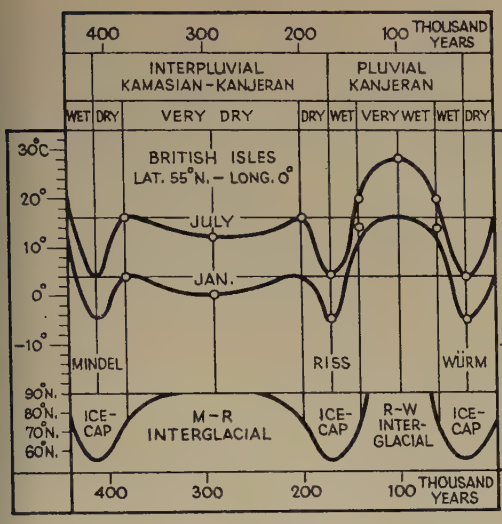


Fig. 3. British Isles.

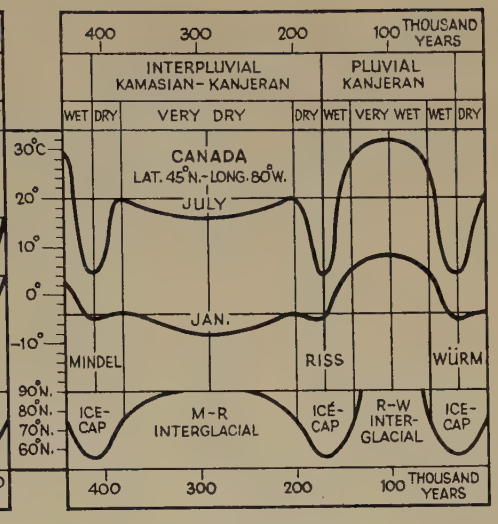


Fig. 4. Canada

Figures 3 and 4. Local temperatures.

relatively recent period. In each of these interglacials the ice-cap was small, probably smaller than at the present day, and the Gulf Stream was in full flow. We may therefore accept that the present-day anomalies existed during both of these interglacials.

With the ice-cap at lat. 75°N, as at present, the Gulf Stream produces in January a large positive anomaly over the NE. Atlantic Ocean. In July the warmer waters of the Stream counteract the formation of the negative anomaly which is usually found over the oceans in the summer months, so that in July there is no anomaly over NE. Atlantic, including the British Isles (Hann-Suring 1926, anomaly charts, Figs. 14 and 15).

The present-day anomalies of the British Isles are January + 15° and July 0°. The zonal temperatures of lat. 55°N during the Mindel-Riss and the Riss-Würm interglacials are contained in columns I, II, IV and V of Table 1. Adding to these the corresponding anomalies we obtain for the interglacial periods the local temperatures given in Table 2.

TABLE 2. Local temperatures (interglacial) British Isles

		Jan.	July
Mindel-Riss interglacial	Beginning	4°C	16°C
	Middle	0°C	12°C
	End	4°C	16°C
Riss-Würm interglacial	Beginning	14°C	20°C
	Middle	16°C	28°C
	End	14°C	20°C

(These temperatures have been entered in Fig. 3 by dots in circles).

(b) *The local temperatures during the glacial periods*

During each glacial period the ice-cap advanced from lat. 75°N to lat. 55°N, then retreated to lat. 75°N and probably disappeared entirely. At the beginning of each glacial period when the ice-cap was at lat. 75°N, as at present, the Gulf Stream was well developed and the conditions of an interglacial period as described in the previous section were still

in existence. With the advance of the ice-cap, the North Atlantic became filled with floating ice and at the maximum of the glaciation when the ice front was on the average along lat. 55°N the whole of the North Sea was full of ice, and pack-ice covered the surface of the Atlantic north of a line from lat. 50°N on the west coast of the British Isles to lat. 40°N on the east coast of North America. In these circumstances local ice-caps formed on the British Isles. The conditions in England were then typical of the margin of all ice-caps with a January temperature of − 5°C and a July temperature of 4°C (see p. 336). These temperatures have been entered in Fig. 3 at the maximum phase of each glacial period. Fig. 3 is now complete and will be used to describe the local climates of the British Isles.

8. The local climates of the British Isles

(*a*) *The local climate : Mindel-Riss interglacial*

During this interglacial there was very little variation in the temperature of the British Isles either in January or July. At the beginning and end of the interglacial (see Fig. 3) the temperature was the same as at the present day. During the interglacial the radiation passed through its minimum, resulting in a fall of 4° in the temperature. Taking the interglacial as a whole, the mean annual temperature was 8°C, that is 2° below the present-day mean temperature. There would be some decrease in precipitation; but not very much. On the whole the most that can be said of the British climate of this interglacial is that it was practically the same as at present with a small decrease of temperature and precipitation in the middle of the interglacial. There would be no variation in fauna and flora to justify the expression ' warm ' as generally used in describing climate on geological evidence.

(*b*) *The local climate : Riss-Würm interglacial*

In the Riss-Würm interglacial the case was quite different. Throughout the whole of this interglacial the January temperature in the British Isles was above 14°C, that is more than 10° higher than at present; while in July for more than half of the interglacial the temperature was above 25°C and at the maximum reached 28°C, 12° above the present July temperature. A climate with a January temperature of 14°C and a July temperature of 28°C (examples : Hong Kong, Jan. 14°C, July 28°C; Cairo, Jan. 12°C, July 29°C) may well be described as ' warm.'

(*c*) *The local climate : glacial periods*

At the maximum phase of all glacial periods the British Isles were surrounded by seas full of floating ice and there were local ice-caps on high land. The January temperature was probably − 5°C and the July temperature + 4°C. An example of such a climate at the present day is along the east coast of Greenland in the latitude of the Arctic Circle.

(*d*) *The geological evidence*

The only positive evidence we can have of past climates is geological. It is therefore necessary to consider the geological evidence for the temperature in England represented by the curves in Fig. 3. This is not the place, nor am I qualified, to discuss the geological evidence for the Pleistocene climate which has been the subject of much difference of opinion amongst both meteorologists and geologists. I shall therefore try to limit my remarks to accepted evidence. I am very much helped in this by the Presidential Address to the Geological Society of London delivered by Prof. W. B. R. King, F.R.S., on 27 April 1955 (King 1955) and shall as far as possible only use information from this Address.

Professor King does not use for the British glaciations the alpine nomenclature ünz, Mindel, Riss and Würm which I have used in keeping with my conviction that the leistocene climates in all parts of the world are interrelated through the variations in olar radiation. I will therefore continue to use the Alpine names with the understanding hat the Mindel-Riss interglacial includes the interglacial beds between the Lowestoft ill and the Gipping Till; and that the Riss-Würm interglacial includes the beds between he Gipping Till and the Hunstanton Boulder Clay.

There are a number of fossils which are frequently referred to in the literature dealing ith the Pleistocene climate as indications of a ' warm ' climate in which warm is not efined, but may generally be taken to mean warmer than the existing summer (presumably uropean) temperatures. Chief amongst these are the Eemian fauna of Holland, the *orbicula fluminalis* and the *Hippopotamus*. Professor King gives cases of all these being ound in England during the Riss-Würm interglacial; but, so far as I can see, not a single ase of one of them being certainly found in the Mindel-Riss. It is true that there has een much discussion about the *Hippo*, relics of which were found near Leeds and elsewhere n beds thought to be earlier than the Gipping Till, therefore in the Mindel-Riss interglacial. rofessor King says (p. 202) ' but I feel that now there are valid reasons . . . . . for placing hem in the Eemian Interglacial [Riss-Würm] . . . . . ; and again (p. 204) ' Moreover here appears to be gradually accumulating evidence that the majority of the deposits ielding *Hippopotamus* in the North and Midlands belong to one interglacial period, the Eemian.' It is too much to expect that all interglacial beds in England have been properly ssigned to their correct period; but there can be no doubt that there is clear evidence or some ' warm ' fauna and flora during the Riss-Würm interglacial in England; the only uestion is how ' warm ' was it ? Luckily there is good evidence. Both *Corbicula* and *Hippopotamus* are living today in the lower Nile where the temperature is January 14°C nd July 26°C, and they are not found in colder climates. These are the theoretical local emperatures we have found for the Riss-Würm interglacial in England shown in Fig. 3 see § 8 (b)).

Turning now to the Mindel-Riss interglacial : we have already seen (§ 8 (a)) that he theoretical temperatures, both summer and winter, which we have found for this nterglacial are practically those of the present day, with no variations during the interglacial eriod sufficiently large to be of geological significance. With reference to this interglacial rof. King refers to Dr. R. G. West's valuable investigation at Hoxon near Cambridge. n this work he studied an almost unbroken sequence of lake deposits which lie between wo boulder clays, the lower of which is recognized as belonging to the Lowestoft glaciation Mindel) and the upper to the Gipping glaciation (Riss). In these deposits, of which a most thorough pollen analysis was made, no indications were found of a climate much warmer than the climate of Hoxon today. Professor King remarks (on p. 198) of his book : It is of considerable importance that the pollen diagram is quite unlike the Eemian diagrams of the Continent, but much more like those of Clacton and the Holstein Inter-glacials ' (two other Mindel-Riss deposits). This is important also for the solar radiation heory as it adds further evidence of a real climate difference between the Mindel-Riss nd the Riss-Würm interglacials.

9. The local climate in Toronto (Canada)

I decided when drafting this paper to prepare a second example of local temperature similar to Fig. 3 for the British Isles for a locality on the same latitude as England, but with a continental instead of a marine climate. I chose for this purpose a locality at lat. 55°N, long. 90°W near the middle of the North-American Continent. I sent a copy of the diagram to Professor Flint asking for his opinion on the agreement between the theoretical temperature and the geological records. He was good enough to send me the following reply. ' Unfortunately we have no interglacial information from the locality you have chosen. The nearest interglacial locality is Toronto, nearly a thousand miles southeast.

There, interglacial beds, probably but by no means certainly, of Sangamon (= Riss-Würm ?) age, include a fauna and flora implying a mean annual temperature similar to but perhaps 2°C higher than, today's, and a rainfall not significantly different from today's. On receipt of this reply I prepared a diagram of local temperature at Toronto similar to Fig. 3; this is reproduced here as Fig. 4.

This diagram was prepared in exactly the same way as Fig. 3 and the results prove interesting. The theoretical Toronto interglacial temperatures, according to Fig. 4, are:

Mindel-Riss interglacial	January − 6°C,	July 18°C,	mean 6°C
Riss-Würm interglacial	January + 6°C,	July 30°C,	mean 18°C
Present day (Isotherm) interglacial	January − 4°C,	July 20°C,	mean 8°C

Thus, according to the theory, if the Toronto beds had been laid down during the Mindel-Riss interglacial the January and July temperatures would have been − 6°C and 18°C respectively, and the precipitation was not significantly different from today. This agrees with Professor Flint's geological evidence; but places the Toronto beds in the Mindel-Riss, instead of in the Riss-Würm, as suggested by him.

Acknowledgments

I wish to express my sincere thanks to Dr. R. C. Sutcliffe, Director of Research, Meteorological Office, London, who read the first draft of this paper and gave me such sound advice that I rewrote it from beginning to end. Whether he will approve the paper in its present form I do not know; but I do know that it is better than it was for his valuable help.

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Appendix

The construction of curves of zonal temperature of all latitudes from the Equator to the North Pole for January and July during a complete cycle of solar radiation.

In Section 3 of the foregoing paper it has been shown how the zonal temperature for any latitude and any glacial or interglacial phase can be derived provided the temperature at the Equator is known for that phase. Unfortunately the temperature at the Equator is not known for any period other than the present day. In view of this fact an attempt

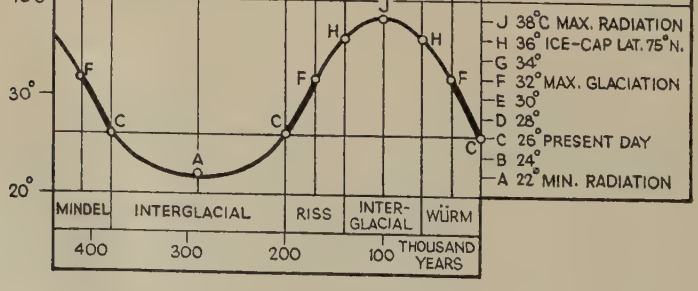

Figure 5. Equatorial temperatures and glacial phases.

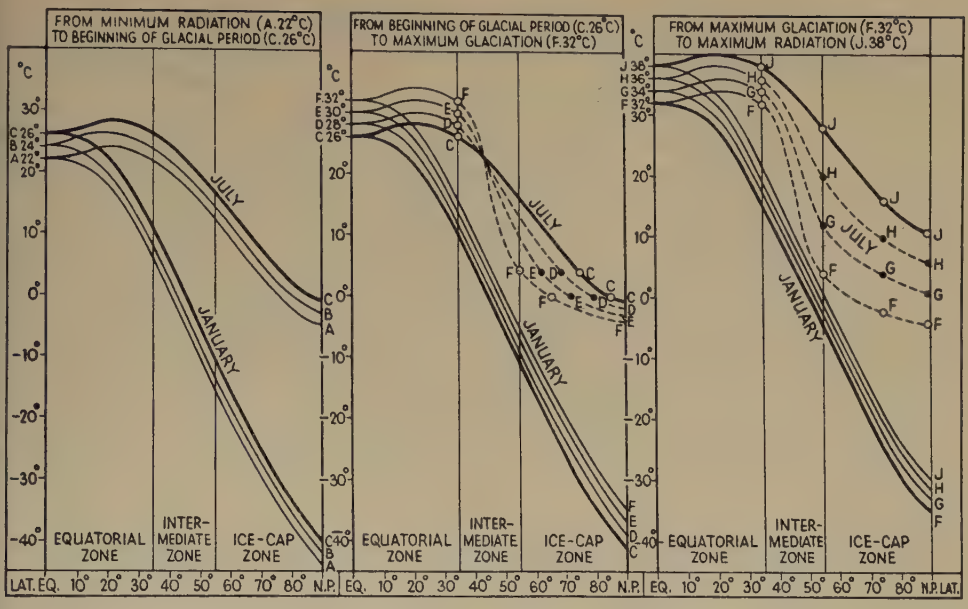
FROM MINIMUM RADIATION (A.22°C) TO BEGINNING OF GLACIAL PERIOD (C.26°C)
FROM BEGINNING OF GLACIAL PERIOD (C.26°C) TO MAXIMUM GLACIATION (F.32°C)
FROM MAXIMUM GLACIATION (F.32°C) TO MAXIMUM RADIATION (J.38°C)
JULY
JANUARY
EQUATORIAL ZONE
INTER-MEDIATE ZONE
ICE-CAP ZONE
LAT. EQ. 10° 20° 30° 40° 50° 60° 70° 80° N.P.
EQ. 10° 20° 30° 40° 50° 60° 70° 80° N.P.
EQ. 10° 20° 30° 40° 50° 60° 70° 80° N.P. LAT.
(a)
(b)
(c)

Figure 6. Zonal temperature and solar radiation.

was made to solve the problem by a graphical process of trial and error in which the relationship between the equatorial temperatures and the zonal temperatures expressed in Lemma I and II were applied to a series of equatorial temperatures varying by steps of 2°C from the present-day value.

In this way the complete set of curves in Fig. 6 (a), (b) and (c) was obtained which fit together so completely that I feel justified in adopting them as an approximate solution of the problem. It is not possible to describe here the stages of the 'trial and error' investigation by which the final solution was obtained, but sufficient information is given to make it possible for anyone interested to examine the evidence on which each curve is based.

Construction of Fig. 5. The equatorial climate

The lay-out of Fig. 5 is similar to that of Fig. 2 previously described. The curve covers the period from the beginning of the Mindel glaciation to the present day, which is a little more than one complete cycle of solar radiation, and as there is no appreciable annual variation of temperature at the Equator, the one curve serves for the individual months and the year as a whole.

The assumption is made that the solar radiation changes so slowly that the atmosphere is always in radiative equilibrium, so that the climate in all parts of the world, including the temperature at the Equator, was always the same for a given value of the solar constant.

For convenience of description the points in Fig. 5 at which the curve crosses the various glacial and interglacial phases are marked by dots in circles to each of which a letter is attached. Each letter therefore specifies both a climatic phase and the equatorial temperature associated with it. On the right of Fig. 5 the letters are shown against the corresponding equatorial temperatures and the chief climatic phases. The glacial and interglacial periods, with a scale of years before the present day, are shown at the bottom of the diagram and the pluvial and interpluvial periods at the top of the diagram.

From actual observations we know only one point on the curve, namely the equatorial temperature at the present day, 26°C, which is represented by the dot marked C at the end

of the curve. A horizontal line is drawn across the diagram from this point to give a base-line from which the variations in the equatorial temperature from the existing temperature can be seen at a glance.

According to the radiation curve in Fig. 1 solar constants equal to that of today occurred also at the beginning of the Riss glaciation and at the end of the Mindel glaciation at each of these phases the equatorial temperature will, by hypothesis, have been the same as today, and these two points also have been shown by a dot in a circle and marked C.

The average latitude of the ice-cap at the maximum of the last three glaciations was 55°N, and the ' trial and error ' investigation showed that with an equatorial temperature of 32°C the zonal temperature of lat. 55°N would be − 5°C in January and 4°C in July, giving a mean annual temperature of 0°C. The investigation showed also that any increase in the solar constant would cause a further rise in the mean temperature and the ice-cap would retreat. This fixes 55°N as the latitude of the margin of the ice-cap at its maximum extent and 32°C as the equatorial temperature at that epoch.

There are three glacial periods shown in Fig. 5 : the Mindel, Riss and Würm and the equatorial temperature at the maximum of each is marked by a dot in a circle at 32°C and labelled F.

We have now six points marked on the diagram; but these alone are not sufficient to complete the curve. The radiation theory, however, gives further help, for according to the theory the equatorial temperature and the solar radiation vary together, therefore their maxima and minima coincide in time. Thus according to Fig. 1 the equatorial temperature will have a minimum in the middle of the Mindel-Riss interglacial and a maximum in the middle of the Riss-Würm interglacial. The ' trial and error ' investigation was continued. Each pair of points on the curve marked C and F were first connected by short lines which are portions of the desired temperature curve : this leaves the two curved portions at the maximum and minimum to be supplied. A few freehand trials soon showed that any reasonable curve would have a maximum somewhere near to 40°C which is 14° above the present-day equatorial temperature. Now this is a very high temperature which might be dangerous to the survival of the present-day fauna and flora, especially as the increased precipitation would involve high humidity. It was clear that the curve could not go above 40°C and I finally adopted the curve, rising to 38°C at the maximum, shown in Fig. 5.

Different considerations apply to the minimum in the Mindel-Riss inter-glacial (the Kamasian-Kanjeran interpluvial). According to the theory, the solar radiation was below its present-day value throughout the Mindel-Riss interglacial (see Fig. 1). The equatorial temperature curve must therefore have been below its present-day value, i.e., below the base-line at 26°C during the whole of the Mindel-Riss interglacial.

Bearing these considerations in mind the curve shown in Fig. 5 was drawn freehand joining up the three portions of the curve already drawn in each glacial period. According to this completed curve the maximum equatorial temperature would be 38°C, which is 12° higher than at the present time. This appears to me to be a high value but not intolerable, especially on high plateaux.

On the other hand, the curve shows an equatorial temperature of 22°C at the minimum of the radiation during the Mindel-Riss interglacial, that is a temperature 4° below the existing equatorial temperature. There is no objection to this temperature with regard to fauna and flora; in fact it would be difficult to distinguish this temperature from the temperature of today on geological evidence.

We now have in Fig. 5 a curve of the temperature at the Equator for a complete cycle of the solar radiation. It is only necessary to apply the appropriate lemmata to these temperatures to obtain the values of the zonal temperatures of all latitudes, in both January and July, and in all glacial and pluvial phases. This has been done in graphical form in Fig. 6 (a), (b) and (c), and in numerical form in Table 1.

It is not proposed to describe step by step how the curves were developed, but Table 3 has been prepared to put on record the lemmata used in building up each portion of the curves.

TABLE 3. METHOD OF CONSTRUCTION OF FIG. 6 (a), (b) AND (c)

I	II	III	IV	V	VI	VII
Equ. Temp. E	ΔE	Phase	Month	Equatorial Zone Eq.-35°N	Latitude Intermediate Zone 35°N-55°N	Ice-cap Zone 55°N-N.P.
			Figure 6 (a)			
22°C	− 4°	A Minimum Radiation Interglacial	Jan.	Lemma I	Lemma I	Lemma I
			July	II (b)	II (c)	I
24°C	− 2°	B Interglacial	Jan.	I	I	I
			July	II (b)	II (c)	I
26°C	0°	C Present day Interglacial	Jan.	I	I	I
			July	II (b)	II (c)	I
			Figure 6 (b)			
28°C	+ 2°	D Glacial	Jan.	Lemma II (a)	Lemma II (a)	Lemma II (a)
			July	II (b)	Interpolate C-F	II (d)
30°C	+ 4°	E Glacial	Jan.	II (a)	II (a)	II (a)
			July	II (b)	Interpolate C-F	Interpolate C-F
32°C	+ 6°	F Maximum Glaciation	Jan.	II (a)	II (a)	II (a)
			July	II (b)	II (d)	II (d)
			Figure 6 (c)			
34°C	+ 8°	G Glacial	Jan.	Lemma II (a)	Lemma II (a)	Lemma II (a)
			July	II (b)	Interpolate F-J	Interpolate F-J
36°C	+ 10°	H End of R-W Interglacial	Jan.	I	I	I
			July	II (b)	Interpolate F-J	Interpolate F-J
38°C	+ 12°	Maximum radiation Interglacial	Jan.	I	I	I
			July	II (b)	I	Interpolate F-J

551.511.6 : 551.556.4 : 53

Deposition of iodine-131 in Northern England in October 1957

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(Manuscript received 23 February 1959)

SUMMARY

Iodine-131 and other radio-elements were emitted from the stack of the Pile No. 1 at Windscale, Cumberland, during the night of 10-11 October 1957. The detailed distribution of the activity deposited on the ground was determined as far south as Preston and Leeds. The deposition of the radio-iodine is attributed mainly to diffusion to the grass, but washout by rain and precipitation from some nearby cooling towers may have increased the rate of fallout in some areas.

1. INTRODUCTION

The accident at the Windscale works of the United Kingdom Atomic Energy Authority occurred on 10 and 11 October 1957. Some of the uranium fuel became white-hot and there was an emission of fission products, especially those which are volatile or have volatile compounds. The course of events has been described in the *Report of the Committee of Investigation*, Penney (1957). Chamberlain and Dunster (1958), Stewart and Crooks (1958), and Dunster, Howells and Templeton (1958) have described the environmental survey carried out after the accident. The purpose of the present paper is to give the isopleths of deposition of activity in detail. A discussion in this *Journal* of these and other features with reference to the meteorological data, follows in the paper by Crabtree.

2. EMISSION OF ACTIVITY

The activity was emitted from the 400 ft stack of the pile, which stands on the west coast of Cumberland (Figs. 1, 2). It is believed that there was serious emission from about 16 GMT on 10 October to 11 GMT on 11 October, with peaks about midnight on 10 October and again during the period 09 hr to 11 hr on 11 October when water was used to put out the fire. During the earlier part of the release, the wind at the height of the stack-top was light and mainly from the SW. With the passage of a cold front over Windscale about 01 hr to 02 hr 11 October, the wind veered to NW. and became stronger.

Neither the total amount of activity in curies nor the variation of the emission with time is known from measurements at the source, but the relative amounts of the various fission products emitted is known and the principal nuclides are listed in Table 1.

TABLE 1. NUCLIDES EMITTED FROM PILE

Nuclide	Boiling point (°C)	Half life	Relative activities emitted
Iodine-131	183	8·0 d.	1.00
Tellurium-132	450*	3·2 d.	0·8
Caesium-137	670	33 y.	0·02

* Temperature at which TeO_2 sublimes.

The physical and chemical nature of the emitted activity is uncertain. Iodine-131 apour is released from irradiated fuel on oxidation of the uranium. The uranium is itially very pure, and the total mass of iodine element formed by fission in the fuel is xtremely small. Probably less than a gram in all was released in the accident, and the olumetric concentration of iodine in the pile stack was of the order of micrograms per ubic metre or less. Nevertheless it is not at all certain that the iodine was all present vapour. Laboratory studies have shown that iodine-131 vapour adsorbs or chemisorbs ery strongly on many metals and oxides, and on filter paper and other substances when he activity is in ' carrier free ' form, that is with little of the stable isotope present, and hat it is not readily desorbed at temperatures up to 400°C (Chamberlain 1953a, hamberlain and Wiffen 1958, Hudswell 1958).

There were fumes of lead and bismuth present in the pile during the accident and dine-131 was found adsorbed on particles of oxide in the stack filters. It is therefore kely that some of the emitted activity was adsorbed on such particles. Some particles f graphite, up to 15 μ in diameter, with adsorbed activity were also found in the neighourhood of the pile after the accident (Chamberlain and Dunster 1958).

Radioautographic studies were made of filter papers through which air had been rawn during the release of activity (Heard 1958). Some particles were distinguished n them, and were found to be mostly formed of graphite with adsorbed activity. However, ot more than 5 per cent of the activity collected near the works was associated with articles of micron size or upwards, and the proportion was even smaller in samples from outhern England. As regards the bulk of the emitted iodine-131 the study of the filter mples leaves open the question whether the activity was in gaseous form or adsorbed n sub-micron particles.

3. Methods of measuring deposition of activity

The methods were of two kinds :

(a) Laboratory analysis of samples of grass, or grass and soil, from known areas of round, and in one instance of the contents of a metal box left exposed to fallout.

(b) Surveys in the field of the gamma dose rate from the deposited activity, using a nsitive radiation monitor originally designed for geological survey.

The sensitivity of the methods under the conditions of the accident is given in Table 2, terms of microcuries of iodine-131 per square metre (μc m^{-2}) of ground. Further etails are given in the references.

TABLE 2. Methods of determining deposition of iodine-131

Method	Sensitivity μc m^{-2}	References
(1) Gamma spectrometry of grass	0·001	Booker 1958
(2) Gamma spectrometry of grass and soil	0·04	Booker 1958
(3) Gamma surveys (ground level)	0·1	Dunster *et al.* 1958 Chamberlain 1958

The derivation from the measurements of the deposition of iodine-131 in $\mu c\ m^{-2}$ c ground took into account the radioactive decay of the activity between the time of emissio and the time of measurement, and also the removal of activity from the herbage to th soil by physical and biological processes. No evidence of removal of iodine by volatiliza tion from the grass was found.

It was necessary to make careful allowance for the contribution of the other activitie besides radio-iodine to the observed dose rates, and to correct for ' background ' radiatio from natural radioactivity in the soil and from fallout of fission products of distant nuclea bomb tests.

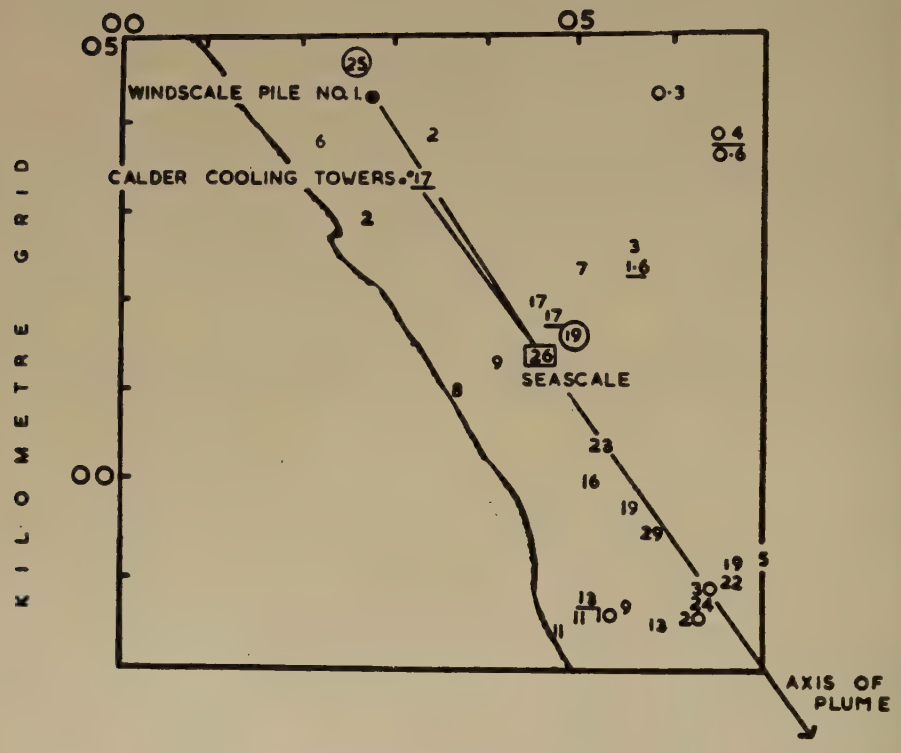

Figure 1. Deposition of iodine-131 near Windscale.

In this and Figs. 2 and 3, the activities of iodine-131 in $\mu c\ m^{-2}$ are indicated thus :

Gamma survey 17. Grass samples 17. Grass and soil samples (19). Pot collection [26].

The degree to which the independent measurements by methods (1), (2) and (3 give consistent results is shown in Figs. 1, 2 and 3. There are a few anomalous result but, in general, the agreement is good. Some of the grass samples show lower result than adjacent gamma readings, but the average of all such comparisons indicates that abou 80 per cent of the iodine-131 activity was initially on the grass (Booker 1958).

Figure 2. Deposition of iodine-131 in west Cumberland and Furness.

4. Isopleths of deposition

Figs. 1, 2 and 3 show deposition measurements at progressively greater distances from Windscale. The 10 μc m^{-2} isopleth in Fig. 2 extends along the coast for 22 km SE. of Windscale, and there is a further small area at this level in the hills of the Furness peninsula. The 1 μc m^{-2} isopleth shows in addition plumes running NE. up the Wasdale, Eskdale and Duddon Valleys.

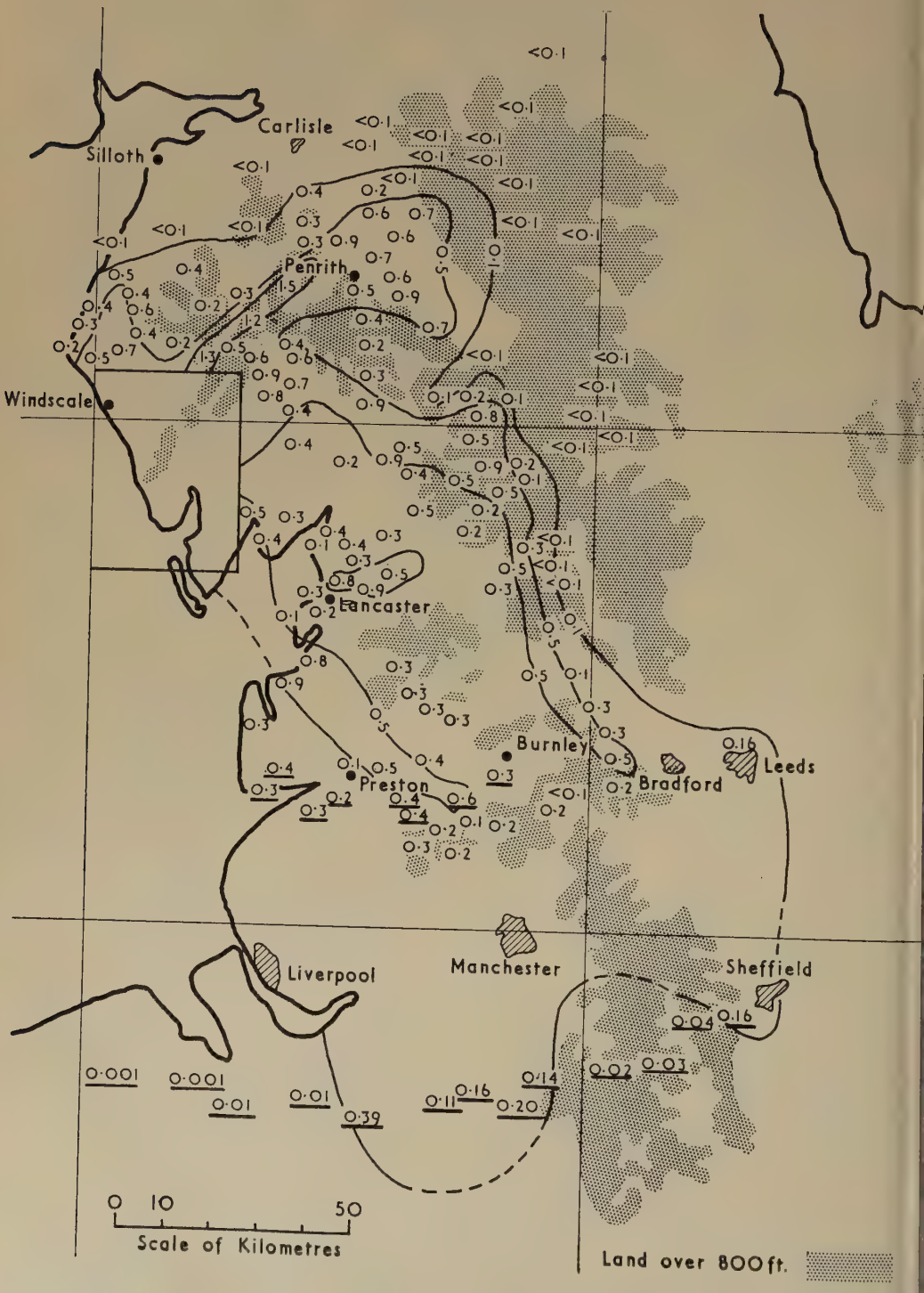

Figure 3. Deposition of iodine-131 in NW. England.

Fig. 3 shows that the activity spread NE. as far as the Pennines, near Penrith, but it did not penetrate to the east of the Pennines to any significant effect, nor did it reach the Scottish border. In Lancashire and Yorkshire two plumes running SE. were found, the 0·5 $\mu c\ m^{-2}$ isopleths reaching to the line Preston to Bradford.

The discussion of these features in terms of the meteorological situations is given in this *Journal* (Crabtree).

The total activity deposited on land in northern England within the 0·1 $\mu c\ m^{-2}$ isopleth can be estimated from Figs. 2 and 3 to be 5,000 curies. In his paper, Crabtree estimates that 22,000 curies passed over the arc Liverpool, Manchester, Leeds. The total emission from Windscale must therefore have been about 27,000 curies.

5. Cross-section of plume running SE. from Windscale

The measurements along the Cumberland coast are sufficiently detailed for the derivation of the cross-sections of the plume at several distances. It must be emphasized that the isopleths of deposition do not necessarily represent the isopleths of air concentration either at one particular time or integrated over the period of emission. This point will be taken up again later.

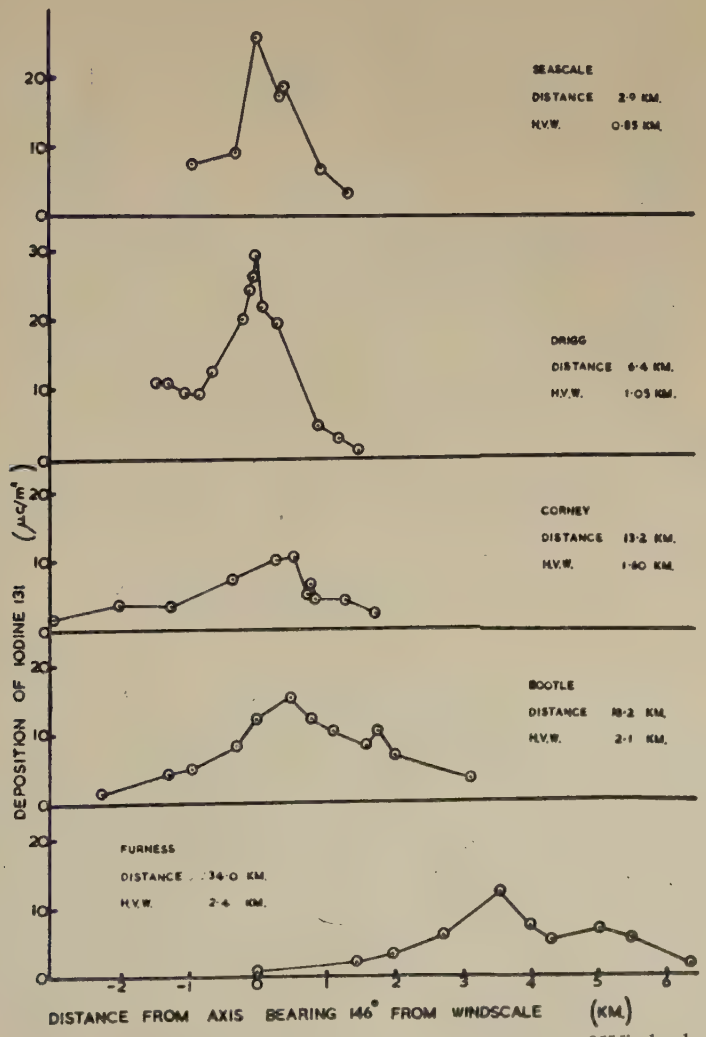

Figure 4. Cross-sections of plume of deposition to the SE. of Windscale.

The cross-sections of the plume at five distances from Windscale are shown in Fig. 4 (The results from gamma survey measurements and from soil or pot collections only have been used). The ordinate in Fig. 4 is the lateral distance from the line bearing 146° from Windscale. The cross-sections at 2·9 km and 6·4 km are truncated to the west by the coast line.

In Fig. 5 the variation of half-value width (H.V.W.)* with distance is plotted on a log-log scale. A tentative value has been inserted for the H.V.W. at 300 km. This is the distance from Windscale of the survey in the South Midlands carried out by Stewart and Crooks (1958). In that survey two peaks were found, one at Wellingborough and the other near Peterborough. If the Wellingborough peak, which lies on the line bearing 140° from Windscale, was caused by the same cloud as that which gave deposition on the Cumberland coast, it can be deduced that the H.V.W. was about 30 km. However the identification of this plume with that along the Cumberland coast is uncertain.

Fig. 5 shows that the H.V.W. of the plume increased approximately as the square root of the distance in the range from 2·9 km to 34 km distance from Windscale.

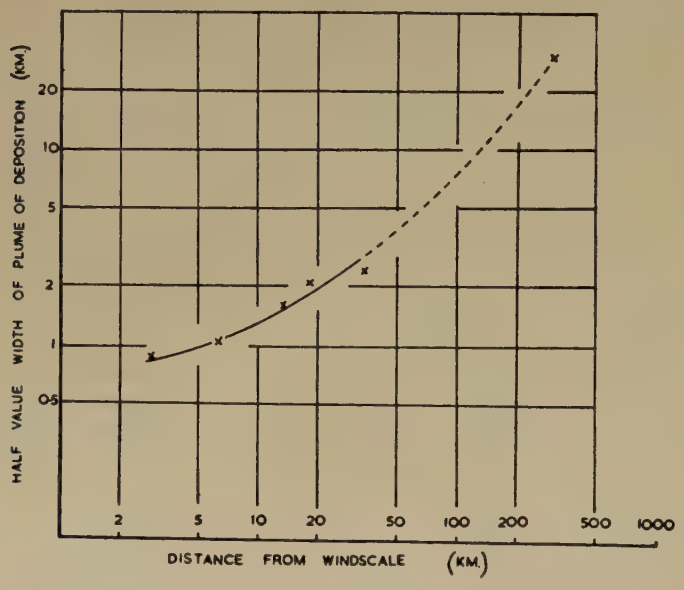

Figure 5. Variation of plume width with distance.

6. Variation of deposition with distance

The variation of the amount of iodine-131 on the ground with distance along the axis of the plume SE. of Windscale is shown in Fig. 6, and the corresponding graph for the deposition of caesium-137 is also shown. The results for caesium-137 were obtained from soil cores taken some months after the accident. Fig. 6 shows that the maximum fallout of iodine-131 occurred at a distance of 6 km from Windscale. There was a second maximum at a distance of 20 km where the cloud passed over the mountain, Black Combe. The stack height is 125 m, but at least during the earlier period of emission, when the wind was light, there was probably considerable thermal rise of the gas from the stack.

* Half value width = width over points where abscissa is half that at the centre of the curve.
= 2·36 × standard deviation for Gaussian distribution.

The deposition of caesium-137 shows a variation with distance similar to that of iodine-131. When the soil samples were taken for caesium-137 analysis it was known that the data on the close-in fallout of iodine-131 were incomplete, and samples were therefore taken at three places on the plume axis between Windscale and Seascale. The first of these was only 0·8 km from the No. 1 pile, and within Calder Works. It gave a fairly high reading of 0·66 μc m^{-2} which is probably partly adventitious in origin. The next two samples at 1·8 km and 2·4 km were much lower and show the existence of a skip distance before the main plume reached ground level. From 2·9 km to 8·4 km the deposition of caesium-137 shows a rather flat maximum corresponding approximately to the maximum of the iodine-131 curve.

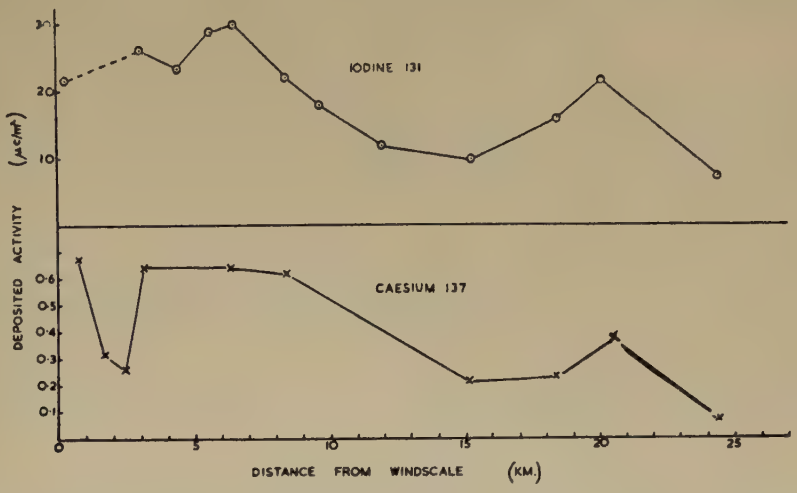

Figure 6. Deposition of iodine-131 and caesium-137 along axis of plume SE. of Windscale.

7. Mechanism of deposition

By combination of measurements of the activity of iodine-131 in air, and on the the ground, an estimate of the velocity of deposition can be made.

$$\text{Velocity of deposition } (v_g) = \frac{\text{Rate of deposition per unit area}}{\text{Volumetric concentration in air}} \quad . \quad . \quad . \quad . \quad (3)$$

$$= \frac{\text{Amount of deposition per unit area}}{\text{Time integral of volumetric concentration } (= \text{dosage})} \quad . \quad . \quad (4)$$

Strictly v_g is a function of the reference height at which the volumetric concentration is measured, but this is not very important in practice for reference heights in the range 1 to 5 m.

For a cloud of large particles the velocity of deposition is the terminal velocity of the particles, but the term may also be used for the deposition of small particles or of gases for which terminal velocity is not the relevant factor.

In Lancashire and Yorkshire a comparison of the deposition results from Fig. 3 and the results of the filter activity measurements of Stewart and Crooks can be made at several locations, as shown in Table 3.

TABLE 3. VELOCITY OF DEPOSITION OF IODINE-131

Location	Deposition of iodine-131 Method	$\mu c\ m^{-2}$	Dosage of iodine-131 $\mu c\ sec\ m^{-3}$		v_g $cm\ sec^{-1}$
Preston	Grass analysis	0·34 0·23	58		0·49
Burnley	Grass analysis	0·34	66		0·51
Sheffield	Grass analysis	0·16	50		0·32
				Mean	0·44*
Preston	Gamma survey	0·14	58		0·24
Leeds	Gamma survey	0·16†	31		0·52
Lancaster	Gamma survey	0·20	64		0·31
				Mean	0·36

* Stewart and Crooks derived $v_g = 0{\cdot}30\ cm\ sec^{-1}$ from the grass measurements at Preston, Burnley and Sheffield. The difference between 0·30 and 0·44 arises from a difference in the method of working back the results from the date of collection of the grass samples (16 October) to the date of the accident. They used the 8-day radioactive half life of iodine-131 whereas the 5-day apparent half life in grass samples as observed by Booker (1958) has been used in this paper.

† The deposition of 0·16 $\mu c\ m^{-2}$ was measured by Spiers (1959) in the garden of a house at Leeds.

A value of v_g can also be deduced from measurements in the Harwell area, and is found to be about a factor 3 less than in Lancashire and Yorkshire (Stewart and Crooks 1958).

There are three possible mechanisms by which the deposition could be caused: (a) terminal velocity of particles; (b) diffusion of activity to the ground; and (c) washout by rain, or by condensation caused artificially.

(a) *Terminal velocity*

The velocity of deposition of 0·3 to 0·4 $cm\ sec^{-1}$ could be accounted for by terminal velocity if the aerosol consisted of graphite particles of about 8 μ diameter or of uranium oxide particles of about 4 μ diameter, but the radioautograph studies make it most improbable that a large part of the activity deposited near the pile was in particles as large as this, and quite certain that this was not the mechanism by which deposition occurred in the Midlands or southern England.

(b) *Diffusion to the ground*

A more likely mechanism is the diffusion of the vapour to the grass surface. Short-range field experiments with iodine-131 vapour have given values of v_g averaging 2·5 $cm\ sec^{-1}$ (Chamberlain and Chadwick 1953), and a similar result has been found in field studies at the Hanford plant in the state of Washington, U.S.A., (Parker 1956). Approximate theoretical estimates can be made of the values to be expected on the assumptions that the adsorption or absorption at the ground is so rapid that the vapour pressure of iodine is zero at the surface and the rate of deposition is controlled by diffusion to the surface (Chamberlain 1953a, Chamberlain and Chadwick 1953). The problem is similar to that of evaporation from a saturated surface into a stream of dry air. There are difficulties in assessing the relative importance of eddy and molecular diffusions. Sheppard (1958, Eq. 13) proposes a formula which when applied to the deposition of iodine vapour to a perfect sink at the ground surface would give

$$v_g = \frac{ku_*}{\log_e (ku_* z_1 D^{-1})}\ cm\ sec^{-1} \quad . \quad . \quad . \quad (5)$$

where u_* is the friction velocity in $cm\ sec^{-1}$, k is Karman's constant, D is the diffusivity

of iodine vapour in air ($= 0{\cdot}08$ cm^2 sec^{-1} at 20°C) and z_1 is the reference height in cm at which the volumetric concentration is measured.

The meteorological conditions over Lancashire and Yorkshire at the time of the passage of the active cloud were such that u_* probably did not exceed 20 cm sec^{-1} even over rough grass. Eq. (5) would then give a theoretical value of v_g of about 0·8 cm sec^{-1}, taking $z_1 = 500$ cm as the height at which air measurements were made. Further south, the activity was present in the air near ground level at the time when the surface wind fell very light after dark on 11 October, and under the inversion conditions which then developed the transport of momentum and of iodine to the ground may have fallen almost to zero.

The rate of deposition and the variation of that rate between the North Midlands and southern England can therefore both be explained on the assumption that diffusion of the vapour was the mechanism of deposition.

(c) *Washout by rain*

Washout by rain may have contributed to the deposition process in some localities. Chamberlain (1953b) has made a theoretical assessment of the possible rate of washout of iodine-131 vapour by raindrops falling through the active cloud. It was assumed that diffusion of vapour to the surface would be the controlling mechanism, the surface being now the surface of a raindrop instead of the surface of grass. The results of Best (1950) were used for the spectrum of raindrop sizes.

Defining Λ as the proportion of the iodine-131 washed out of the air per second, it was calculated that Λ increased from 1×10^{-4} to 2×10^{-4} sec^{-1} as the rate of rainfall increased from 1 to 5 mm hr^{-1}*.

The relation between the washout coefficient and the rate of deposition depends on the height of radioactive cloud through which the rain falls. If the iodine-131 is limited to the first few hundred metres of the air, and the rain is falling through it from rain clouds above, Λ is approximately independent of height. The ratio of the rate of deposition by washout I_w to the rate by direct deposition I_d, can then be shown to be

$$\frac{I_w}{I_d} = \frac{\Lambda H}{v_g} \quad . \quad . \quad . \quad . \quad . \quad (6)$$

when H is the effective height of the cloud (i.e., that height which, when multiplied by the air concentration at ground level, gives the integral of the air concentration with height).

From theoretical consideration Λ is about 10^{-4} sec^{-1} in light rain and from both theoretical and experimental work v_g is about 10^{-2} m sec^{-1}. It follows that I_w can be expected to be comparable with I_d if H is of the order of 100 m. By the time the plume from a 125 m stack has diffused to ground level H is about 300 m, and as the cloud travels downwind it is likely to increase at least to 1,000 m or 2,000 m. It follows that if light rain occurred at a time when the cloud of vapour was passing, the deposition from that cloud would be considerably increased.

There was patchy rainfall in north-west England during the rainfall day 09 hr 10 October to 09 hr 11 October. The highest totals (8-10 mm) were recorded in the Furness peninsula, but from records of a rainfall-recording gauge at Barrow it is known that most of this fall occurred between 17 hr and 19 hr on 10 October, about 6 hr before the passage of the front and before the active cloud had reached the area. Traces of rain between 01 hr and 02 hr on 11 October, about the time of passage of the front, were, however, recorded at Silloth and it is likely that any rainfall associated with the front would be heavier in the hills. The front became very weak as it passed south and there was no rain associated with it in the Midlands and southern England.

* Λ is not a linear function of rainfall rate because light rain consists of smaller drops than heavy rain, and the smaller the drops the greater the surface area per unit mass and the greater also the rate of diffusion to the drop.

The conclusions are that diffusion of iodine-131 as a gas to the grass could account for the rate of deposition over England and Wales as a whole. If, as seems likely, the activity was emitted partly as a gas and partly adsorbed on sub-micron particles, the former fraction would be expected to be deposited more rapidly, since it has much higher molecular diffusivity. If it is assumed further that the caesium-137 activity was more readily adsorbed on particles, the preferential deposition of iodine-131 relative to caesium-137 would be expected, and this would explain the finding of Stewart and Crooks (1958) that the iodine-131/caesium-137 ratio on filter-paper samples decreased with distance from Windscale.

The role of rainfall in assisting deposition could theoretically be great, but the information is not sufficient to say whether or not some of the fallout in north-west England can be attributed to it.

8. Causation of narrow band of deposition along Cumberland coast

Three factors may have caused the narrow plume of deposition along the coast, shown by the 10 μc m^{-2} isopleth in Fig. 2, as follows :

(a) A greatly increased rate of emission for a period of the order of an hour, some time after the N. to NW. wind became established early on 11 October.

(b) The existence during a period of factors enhancing deposition, for example, rain or turbulent air.

(c) The effect of the entrainment of the active cloud from the Windscale pile with the water vapour and droplet cloud from the cooling towers at the Calder A nuclear power station.

The explanation (a) is feasible if the period of high emission causing the plume of fallout is identified with the time about midnight on 10 October, when high activity measurements were made in the Calder area (Penney 1957). At the later period of enhanced emission from 09 hr to 1030 hr on 11 October, the cloud is known from local observation to have been carried about due S. from Windscale and out to sea, and it is also known from local measurements (Leslie 1958) that the deposition of 26 μc m^{-2} in Seascale village, shown in Fig. 1 nearly all took place before 0830 hr on 11 October.

It is possible that factors (a) and (b) above, were both operative immediately after the passage of the front in the early hours of 11 October.

The cooling towers at the Calder A power station, which was in operation at the time of the Windscale accident, are 300 ft high and are situated 1 km SSE. of Windscale Pile No. 1. In Fig. 1 are shown the relative positions of the pile, the cooling towers, and the axis of the plume of deposition. The difference in bearing of the first two from the point of greatest deposition in Seascale is only 2°. Such a difference in the direction of the plumes from the pile and the cooling towers could be accounted for in magnitude and in direction by the probable veer of wind direction from 300 ft to 400 ft altitude. It is therefore probable that the cloud of iodine-131 from Windscale Pile No. 1, would be entrained over Seascale with the cloud of vapour and droplets from the cooling towers, which would be persistent in the conditions of high relative humidity of the night of 10 to 11 October.

The close alignment of the axis of the deposition pattern with the line joining the Windscale pile and the Calder cooling towers may be a coincidence, but it seems very probable that spray or condensate from Calder enhanced the deposition. There is some evidence also of an effect on the reverse bearing, that is the NNW. of the pile, where the two plumes would also mingle if the wind direction was suitable. The 1 μc m^{-2} isopleth in Fig. 2 shows a bulge in this direction.

The fact that the maximum deposition to the SE. of the pile was found in the Seascale-Drigg area 3 km to 6 km away is not necessarily inconsistent with the present hypothesis. The pile stack is 125 ft higher than the cooling towers and the gas temperature also was higher in the pile stack, so that the plumes would travel a considerable distance before their vertical spread caused them to intermingle.

Acknowledgments

I am indebted to Mr. H. Howells, Manager Health Physics, and other members of the staff at Windscale for access to measurements made there and to Mr. D. V. Booker who made the analyses of activity in grass and soil by gamma spectrometry.

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The travel and diffusion of the radioactive material emitted during the Windscale accident

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(Manuscript received 9 June 1959)

SUMMARY

The reactor accident at Windscale on 10-11 October 1957 resulted in the release of radioactive fission products into the atmosphere over a period of several hours. Measurements were made of the fission products, particularly iodine-131, collected on filters exposed for the study of atmospheric pollution in many places in the United Kingdom and Europe : measurements were also made of the iodine-131 deposited on grass. These showed that the effluent had been widely dispersed, but that activities over England were about an order of magnitude greater than those found over the neighbouring areas of the continent. Meteorological trajectory analysis showed that the main portion of the cloud had remained over England until 15 October, and that the activities over continental Europe were due to the eastern edge of the cloud.

Meteorological data were used together with the air concentration data to deduce that about 22,000 curies of iodine-131 were present in the cloud as it passed over south Lancashire and Yorkshire.

1. INTRODUCTION

As a result of the accident in the No. 1 Pile at Windscale on 10 to 11 October 1957, a quantity of radioactive material was released into the atmosphere from the 120 m reactor stack. Measurements were made, during the course of the release, of the total activity retained by the filter at the top of the stack and of the activity in the air near ground-level at a number of points on the site. These measurements showed that the emission was substantially confined to the period 12 GMT 10 October to 12 GMT 11 October, with probable peaks in the rate of discharge around 24 hr 10 October and 09 hr 11 October : they did not, however, yield any detailed information on the variation of the rate of emission with time, nor did they enable an estimate to be made of the total amount of radioactivity released.

The emission occurred during a period when meteorological conditions over the country were changing rapidly, and the effluent was, in consequence, widely dispersed. Measurements of the fission products, particularly iodine-131, collected on filters exposed as a routine in many industrial areas by local authorities and other bodies, have been briefly described by Stewart and Crooks (1958) : from a study of these data the travel of the radioactive cloud can be followed as far as south and east Europe and Scandinavia. The deposition of iodine-131 on grass in the United Kingdom is described in the immediately preceding paper by Chamberlain (1959).

The purpose of the present paper is to discuss the history of the radioactive cloud by making use of the activity observations and meteorological data, including trajectory analysis, and to give an estimate of the amount of iodine-131 released into the atmosphere during the course of the accident.

2. THE ACTIVITY DATA

(*a*) *Iodine deposition data*

The results of a survey of the iodine-131 deposited over northern England are described in the preceding paper. The deposition data provide a valuable aid in determining the behaviour of the radioactive cloud over Cumberland, Westmorland and north Yorkshire where no air concentration data are available.

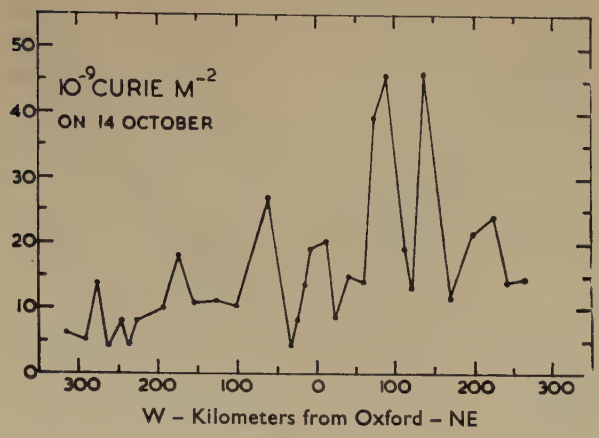


Figure 1. Survey of iodine-131 on grass.

Data on the deposition of iodine-131 on grass along a line from south-west Wales, through Oxford, to East Anglia have been presented by Stewart and Crooks (1958): the results are shown in Fig. 1. The most striking feature is the length of the arc along which deposition occurred, this length, about 600 km, being nearly twice the distance of the arc from Windscale. Other noteworthy features are the two narrow peaks at about 90 km and 140 km north-east of Oxford, with relatively low activity between, and the smaller peaks near Oxford and 60 km to the west.

(*b*) *Air concentration data*

The results of the measurement of iodine-131, caesium-137 and polonium-210 collected on filters, exposed for the study of atmospheric pollution, are available for many industrial areas in the United Kingdom and Europe. Many of the samples from Europe were obtained through the chairman of the International Geophysical Year Advisory Committee on Nuclear Radiation, Professor W. Bleeker; those from the United Kingdom were obtained from local authorities and from the Fuel Research Establishment. The filters are normally changed every 24 hr (48 hr at week-ends in many cases), so that some idea can be obtained of the travel and spread of the radioactive cloud, albeit on a rather coarse time scale. This is illustrated in Figs. 2 to 5, where the iodine-131 activity distributions are plotted for three 24-hr periods and one 48-hr period: the data have been corrected for decay to a common date, 10 October.

Moving strip filters were exposed at the Kodak Works, Harrow (near London), at Antwerp and Mol in Belgium, and at a number of stations in Germany. The total beta activity on each filter has been measured as a function of time, enabling the time of arrival of the cloud at each station to be determined. The results for a few stations are shown in Fig. 6, the scale being in arbitrary units and varying from station to station: in Fig. 3 these stations are identified by their initial letters.

Fig. 2 shows the distribution of iodine-131 activity in air during the 24-hr period commencing at approximately 09 hr 10 October. The pattern is very similar to the deposition pattern: there are two main plumes, one over west Yorkshire and the other over Lancashire. Fig. 3 shows the iodine-131 activity for the next 24-hr period: a region of high activity extends from south Lancashire to London, with fairly high activities along the south coast of England. The isopleths are drawn as a broken line over areas where observations were scarce or entirely lacking. The active cloud has spread to Belgium, the extreme west of Germany, and to south Holland, but the highest activity in this portion of the cloud is only about six per cent of the London value. The Harrow filter

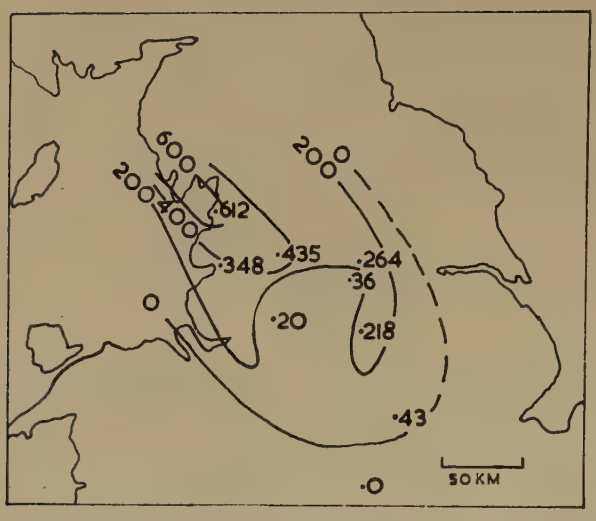


Figure 2. Iodine-131 activity in air (10^{-12} curie m^{-3}) 0900 GMT 10 October to 0900 GMT 11 October.

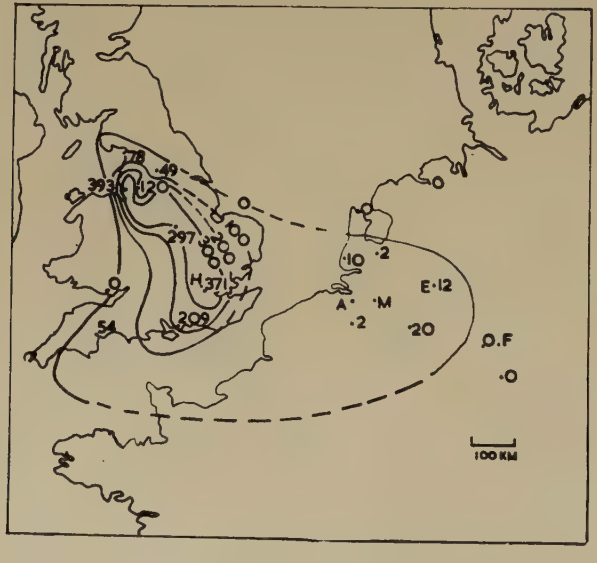


Figure 3. Iodine-131 activity in air (10^{-12} curie m^{-3}) 0900 GMT 11 October to 0900 GMT 12 October.

showed that the cloud first reached the London area shortly before 18 hr 11 October : the peak activity occurred at 02 hr on the next day, and the activity gradually fell to 10 per cent of its peak value by mid-day. The activity first reached Belgium and London almost simultaneously, and reached Frankfurt about a day later.

The iodine concentrations of Fig. 4 are the average value for the 48-hr period beginning about 09 hr GMT 12 October. The concentrations over England have decreased considerably, but the cloud is still quite compact and has moved only very slightly from its position in Fig. 3. The peak activities are found in two areas, one over south central England, the other over the north Midlands. The activities over the continent have, in contrast, shown only a slight decrease from those of the previous period : those filters which were changed on the morning of 13 October show a slow northward extension of the cloud over Holland.

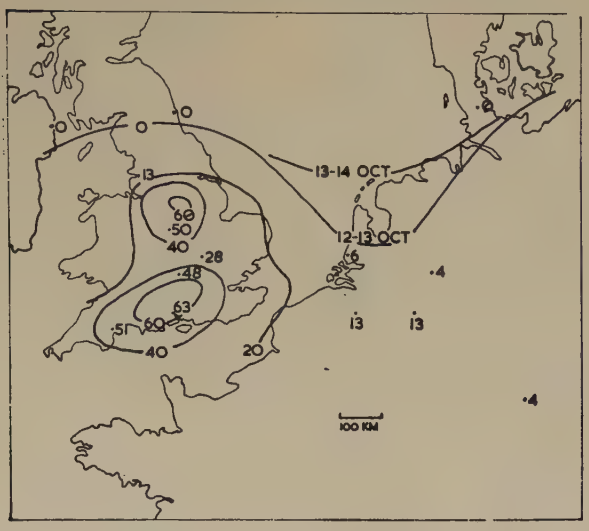


Figure 4. Iodine-131 activity in air (10^{-12} curie m^{-3}) 0900 GMT 12 October to 0900 GMT 14 October.

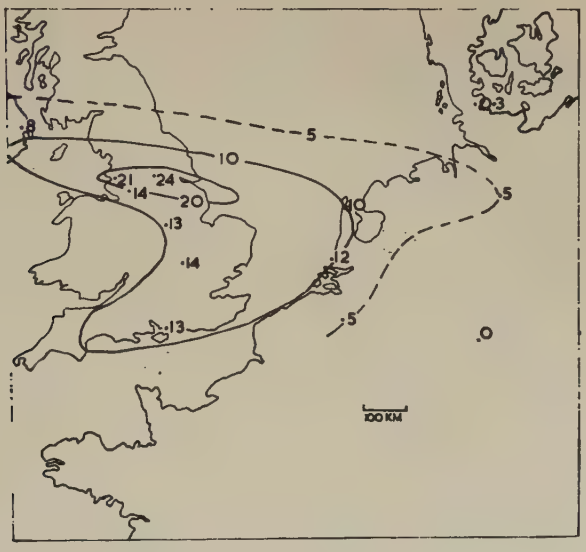


Figure 5. Iodine-131 activity in air (10^{-12} curie m^{-3}) 0900 GMT 14 October to 0900 GMT 15 October.

By 14-15 October (Fig. 5) the activities have again decreased over England, and the cloud has extended a little further north and west. Again, activities over the continent have decreased only slightly. By the 15-16 October, the cloud had moved northeast and reached southern Scandinavia.

3. Meteorological data

(a) *The synoptic situation*

Figs. 7 (a) to 7 (f), show the general synoptic situation over the British Isles at various times during the period of the emission and the following 24 hours. At 12 hr 10 October a ridge of high pressure extended across England from an anticyclone over Europe. A cold front lay over central Scotland and Northern Ireland, with low pressure to the north,

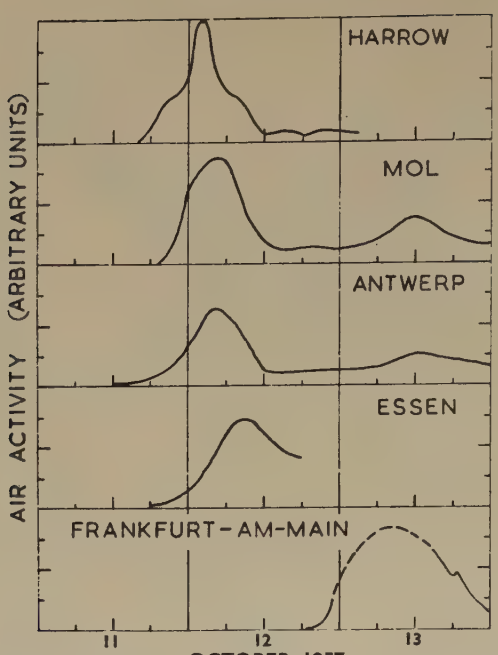


Figure 6. The variation of air activity at a number of stations.
(The location of each station is indicated in Fig. 3 by its initial letter).

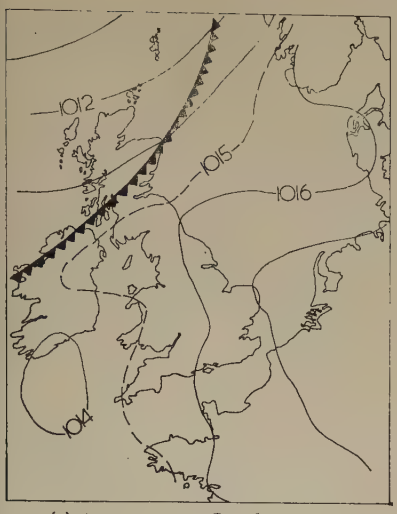
giving mainly light S. to SW. winds over north-west England. A small area of low pressure lay off south-west England : by 18 hr this had moved to south Lancashire, producing light E. or NE. surface winds over north-west England for a few hours : its influence was felt only in the lowest layers of the atmosphere, the wind at 900 m remaining SW'ly. By 24 hr, this low-pressure area had weakened and moved E., and the SW'ly airflow had again become established.

The cold front moved SE. during the day and crossed Windscale early on 11 October, probably about 01 to 02 hr. Conditions were cloudy for some distance ahead of the front, and rain, mainly light, was reported from many places in the Lake District and Pennines. At the passage of the front, the winds quickly veered to a N'ly direction, and gradually backed to NW. during the morning. The wind direction was fairly constant with height in the lowest 1000 m, but during the first few hours after the passage of the front the surface winds remained light and the speed increased rapidly with height. The front continued SE., clearing the Kent coast by 19 hr 11 October, and producing little, if any, rain south or east of the Pennines. By the time the front had crossed the English Channel, it had become very weak and difficult to locate.

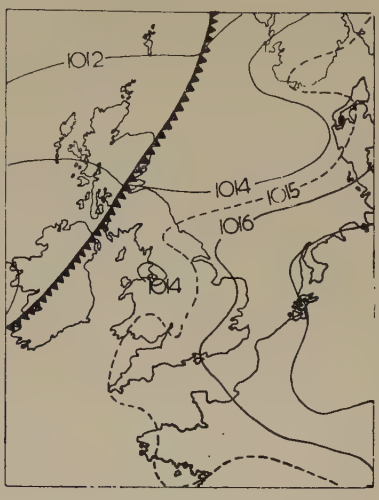
The ridge of high pressure behind the front intensified rapidly as it moved across England, giving east winds over the extreme south of the country. By 12 hr 12 October, the synoptic situation was similar to that obtaining two days earlier. Pressure remained high and winds over England were light and variable until late on 14 October, when an approaching frontal system brought increasing SW. winds to England and north-west Europe.

(*b*) *Analysis of the travel and spread of the cloud*

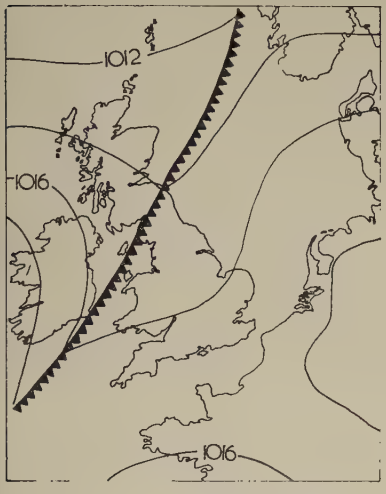
The main features of the deposition and air concentration data for northern England can now be explained qualitatively from a consideration of the synoptic data. Material released into the atmosphere before the passage of the cold front would be carried mainly NE. by the SW. wind which blew for most of the period : it would then be swept SE. as the front approached, producing the plume of deposition and the high air concentrations



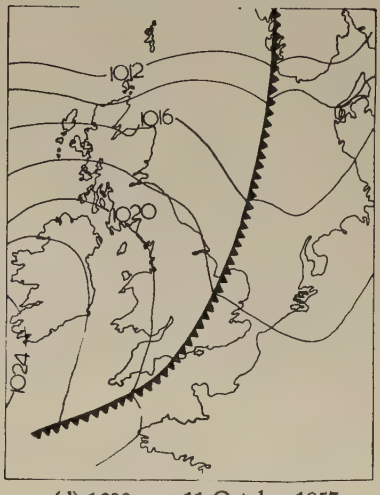
(a) 1200 GMT 10 October 1957



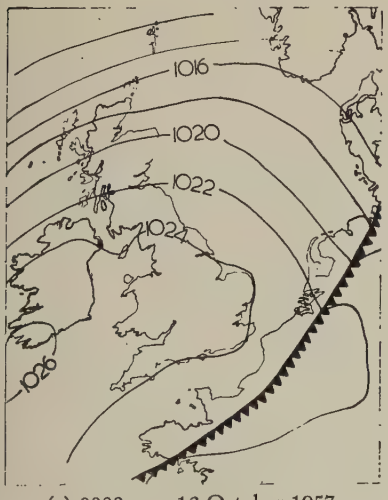
(b) 1800 GMT 10 October 1957



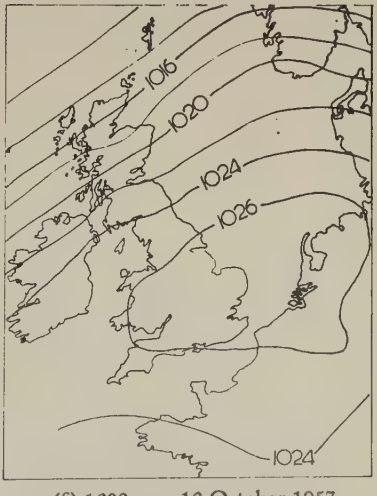
(c) 0000 GMT 11 October 1957



(d) 1200 GMT 11 October 1957



(e) 0000 GMT 12 October 1957



(f) 1200 GMT 12 October 1957

Figure 7. Surface synoptic charts.

over the West Riding of Yorkshire. The western plume of deposition and the high a concentrations over Lancashire were due to material released at about the time of, an after, the passage of the cold front across Windscale.

Some features of the local activity observations discussed by Chamberlain cannot b explained fully with the limited meteorological data available at Windscale. The airflo during the afternoon of 10 October was reasonably straightforward, but the formatio of a temperature inversion in the lowest layers of the atmosphere in the evening seem to have complicated matters. At about 18 hr the surface and gradient winds were E' or N'ly, and the plume slowly drifted out to sea. During the last few hours of the da the SW. gradient wind became re-established, but the surface wind at Windscale sti came from the N. or NE. It is probable that this wind was due to a land breeze or katabat effect, which may not have extended to the top of the 120 m stack. Under these condition the plume would move NE., and as it diffused downwards, be caught in the down-slop wind and return to the coast.

Estimates were made of the travel of the cloud by using meteorological informatic to trace the paths (trajectories) followed by airborne material released at intervals through out the period of emission. A smooth line drawn through the end-points of the traje tories at a given time gives the ' plume ' of emitted material, i.e. the axis of the radioactiv cloud, at that time. Various methods of drawing the trajectories were tried, and th resultant plumes compared with the air activity data. It was found that the use of geo strophic or gradient winds based upon mean sea level isobars was unsatisfactory becaus of large ageostrophic wind components in the rapidly developing anticyclone behind th front. On the other hand, the actual surface winds were generally too light to explai the observed movement of the radioactive cloud. Eventually, the most realistic resul were obtained by drawing streamlines to the observed surface winds, smoothing out loc effects to some extent in the process, and increasing the observed speeds by 25-50 per cen The winds obtained in this way are probably not very different from the actual winds 150-300 m, and very similar trajectories were obtained by using the few 900 m rada wind observations available.

The results of the trajectory analysis are shown in Fig. 8, which depicts the plum of emitted material at 6-hr intervals. The cloud will in practice have diffused parall

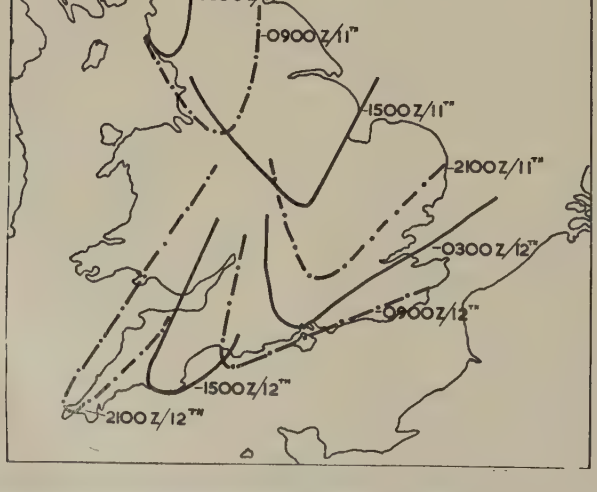


Figure 8. Position of the plume of emitted material at 6-hr intervals, based on meteorological analysis.

o, and at right angles to, the axis. It was assumed that the emission occurred throughout he period 12 hr 10 October to 12 hr 11 October : emission in the first twelve hours f this period is represented by the E. or SE. half of the plume, that in the second half f the period by the W. or NW. part of the plume, the bend being a result of the change f wind direction associated with the passage of the cold front.

The NE. movement of the early emission is confirmed by the trajectory analysis, which uggests, however, that the deposition data would be better explained if serious discharge ad not commenced until about 15-16 hr 10 October : this portion of the cloud probably ave rise to the deposition peak 200-220 km NE. of Oxford (Fig. 1), subsequently passing ver the extreme SE. of England and then moving W. along the south coast. Part of this ctivity diffused across the southern North Sea to Belgium and Holland : if the main loud had crossed the continent, the activities there would have been a factor of 5 or 10 igher than those actually observed, in spite of the fact that the sea temperatures were igh enough to cause appreciable convection. The data imply that the half-width of he cloud was about 200 km after a distance of travel of 600 km.

The emission occurring around the time of the passage of the cold front across Windscale moved SSE., probably giving rise to the deposition peaks at 90 km and 140 km NE. of Oxford. The high air concentrations found in the London area were also due to his portion of the cloud, which, after passing London, drifted slowly W. over southern England. The calculated time of arrival of the plume at Harrow agrees, within a few ours, with that given by the moving strip filter exposed at the Kodak factory.

The material emitted during the last few hours of the release moved SSE. across Lancashire to the Midlands, where it remained, moving only slowly during the period for which the trajectories were drawn. After 21 hr 12 October, the winds were so light and variable that tracking was of little use : little movement took place until 14-15 October when stronger SW. winds reached north-west Europe and carried the cloud NE. to reach Scandinavia by 15-16 October.

A comparison of the observed patterns of activity with the calculated plume positions supports the view that the emission was practically confined to the period 15 hr 10 October to 12 hr 11 October, and that the rate of discharge varied quite considerably with time, probable peaks in the rate occurring at about 15 hr and 24 hr 10 October and 09 hr 11 October.

4. Estimation of the amount of iodine-131 released into the atmosphere

The air activity data for Lancashire and Yorkshire have been used, together with meteorological data, to estimate the total amount of airborne iodine-131 crossing a line from Liverpool to Flamborough Head. This line is approximately perpendicular to the movement of the radioactive cloud. The air concentration data for stations within 30 km of the line are used to deduce that the mean iodine-131 activity-in surface air along that line for the period 09 hr 10 – 09 hr 12 October, was 250 $\mu\mu$c m^{-3}, corresponding to a dosage of 43 μc sec m^{-3}, and that the width of the cloud was 12×10^4 m. It is inferred from the filter data and trajectories that the cloud crossed the Yorkshire portion of the line at about 09 hr and the Lancashire portion between 06 hr and 15 hr 11 October, and further that none of the activity was associated with a second passage of the cloud across the area. The average wind speed during the passage of the cloud was about 5 m sec^{-1}. Upper air temperatures at Liverpool, Aldergrove and Hemsby indicated that the air ahead of the front was mainly stable, and that the air behind the front was neutral or slightly unstable up to about 1,500 m with an inversion above. It is very unlikely that the concentration was uniform throughout this depth : it is assumed that the concentration follows a Gaussian distribution and that the height of the cloud (to where the concentration has fallen off to 0·1 of its ground-level value) was equal to the height of the layer below the inversion, giving a ratio of the average concentration throughout the depth to the surface concentration equal to 0·56.

The integral of iodine-131 crossing the line is then given by :

$$0{\cdot}56 \times 43\ \mu c\ \text{sec}\ m^{-3} \times 12 \times 10^4\ m \times 1500\ m \times 5\ m\ \text{sec}^{-1}$$

$$\simeq 22{,}000\ \text{curies iodine-131}$$

The greatest source of error in the calculation lies in the assumption regarding the vertica mixing : the above estimate could be too high or too low by a factor of up to 1·5. I spite of its roughness, the estimate is of value, as no independent assessment can be mad from the available data.

Acknowledgments

Acknowledgment is due to the Director-General of the Meteorological Office fc permission to publish this paper, to Mr. A. E. Parker for assistance in some of th trajectory work, and to Mr. L. C. Cox for preparation of the diagrams.

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551.506.3

An Irish weather diary of 1711-1725

By F. E. DIXON
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(Manuscript received 13 February 1959; in revised form 19 June 1959)

Summary

The oldest extant weather diary for Ireland is critically examined and the results presented in a series of tables. Some comparisons are made with modern data.

1. Introduction

This paper is based on a study of a manuscript weather diary in the possession of the Royal Society, to whom the observer presented it in 1726. It has been cited by Manley (1953), but no comprehensive account of the diary, or summary of its contents, has been published hitherto.

There were earlier Irish weather records, notably one at Trinity College, Dublin, in the 1680's (Gilbert 1859), but no trace of them can now be found.

The first sections of this paper discuss the diarist and the diary in general. Later sections summarize the particular meteorological elements in turn and compare the results with some modern data.

2. The diarist

Very little can be ascertained about Thomas Neve, the writer of the diary. The *Dictionary of National Biography* mentions several prominent bearers of the name, such as Jeffery Neve (astrologer), 1579–1654, and the Rev. Timothy Neve, 1694–1757, but there is no evidence linking either with Thomas. The Dublin Genealogical Office has a note of a will of a Thomas Neve, written in 1739 and proved in 1749. This will mentioned two daughters and no sons : the diarist mentions his household of 12, and I think it likely that the Thomas who wrote the will was a son of Thomas, the diarist.

The diary discloses that he had earlier lived in England, where he had sat up many nights ' learning the constellations by the Coelestial Globe and observing the Appulses of the Moon with fixt Stars and the Planets, with a 11 ft Telescope and Micrometer.' He possessed a ' Triangular Quadrant ' and a ' good Needle of Surveying Instrument ' which he used for determining azimuths during auroral displays. He owned two watches, but in November 1720 both were away for repairs. He also had a ' Royal Pendulum Clock ' and was able to count seconds by the noise of ' the pads on the swing-wheel.' His reading may be judged from two quotations from the diary : on 19 November 1720, a little after midnight, he was lying in bed reading *Whiston*'s *Astronomical Lectures*. The bright lights in the sky which then caught his attention reminded him of the ' draught of a Comett called by Mr. Lyburn in his *Cursus Mathematicus Dimericatus*.'

Neve was not a member of the Royal Society and he communicated with it through intermediaries, Derham and Pyke. Two of his contributions were published in the *Philosophical Transactions* and were reprinted by Boate and Molyneux (1726).

Neve's Irish home was at Ballyneilmore (often referred to as Ballyneil) in the parish of Artrea, Barony of Loughinsholin, and County Derry. This is lat. 54° 42′N, long. 6° 34′W, about 1½ mi WSW. from the shore of Lough Neagh. Fig. 1 shows the location and marks the principal places mentioned in the diary.

Ballyneilmore and its neighbour Ballyneilbeg were among lands settled by the Salters Company in 1613, assigned by the Company to a Captain Whistler in 1627. In 1752 the house is described as a good farm house with very good offices, but most of the 354 acres were rocks, whins, moss and bog, yielding an annual rental of only £27. Subdivision of holdings and increases in rents raised this to £337 in 1841 (Salters Co. 1842).

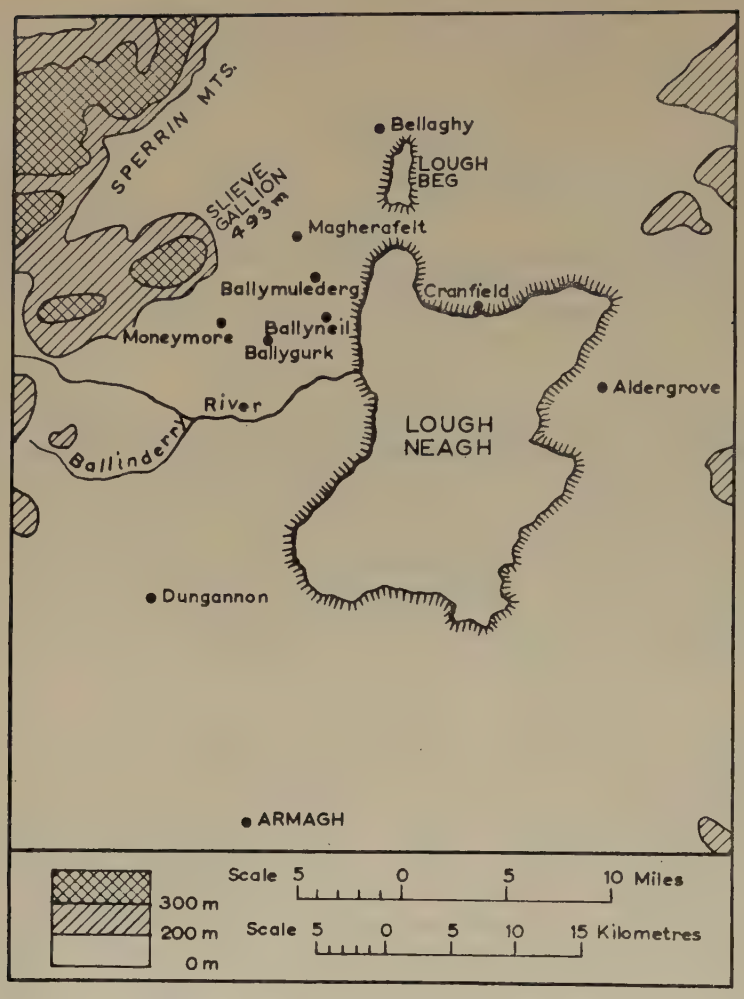

Figure 1. Map of ground around Ballyneil.

3. The diary

The writing is clearly legible and the chief uncertainties in decipherment are occasioned by the use of abbreviations. e.g. ' Fr ' may be frost or fair : ' *Idem* ' following an entry of ' cloudy and rain ' may mean only ' cloudy.'

The diary is not the original register, but a copy, made in 1725 and 1726 for sending to the Royal Society. The copying was not checked. Some gross errors are apparent, and it is likely that there are others less serious. The most extensive slip was that the weather and wind data for the period 10 November to 28 February 1712 ' Old Style ' calendar (O.S.) were repeated for the same period in the following year. It was not possible to determine which were the correct entries, and for the purposes of my summaries they have been accepted as valid for both years. Another difficult case is that of August 1719, which is credited with frost on five days and snow on three. This is improbable, and inspection of the diary suggests that the data for May and August have been interchanged. This assumption eliminates discontinuities of wind and weather at the beginning and end of each month.

Repetition has in one instance been helpful by confirming a translation; 17 February 1712 has the entry ' Album gelu ' and 17 February 1713 has ' White frost.' Elsewhere the words ' frost ' and ' gelu ' were used indiscriminately.

Initially the diary comprised two entries per day, the earlier being at '8' in winter ut '6' in summer. The other hour was '23,' a use of the 24-hr clock – not usual in he 18th century. Occasionally an entry for '12' or '13' was inserted and after 1714 his third daily observation became a regular practice. After 1717 the late observation ecame '22' and towards the end of the diary '5' became more frequent than '6' for he morning observation hour in summer.

Each entry of the weather was normally compressed into not more than three words, .g. 'cloudy,' 'overcast, sm. rain,' 'hazey Gelu dura.' Wind direction and force was ntered each time, as was rain whenever it was measured. Readings of temperature and ressure commenced only in 1722. It is a coincidence that Neve's instruments were nstalled immediately after Rye stopped using a similar pair near Cork (Rye 1734). The cales were, however, different, and Neve probably bought his direct from the maker.

On many pages additional notes are squeezed into spaces above or below the main liary, and from July 1722 more details are reported on separate pages with headings such s 'Memoirs Relating to the Weather, Barometer, Thermometer, etc. in July 1722 More t Large than the Page of the Weather of the Month will contain.' Many of these Memoirs re trivial, describing such happenings as clouds at different levels with different motions, iving the times of showers, mentioning cloud caps on mountains, etc. Neve also noted vith surprise some disagreements between the legend on his barometer and the actual veather : e.g., 'I cou'd not but wonder ye Night yt ye barometer was fallen no Lower ltho' we had a deal of Rain ys Evening and is still Rainey hard and a prety strong Wind' 30 Dec. 1722).

Pages 177–194 are filled with verbose accounts of some varied phenomena. The most notable meteorologically are auroral displays, and other topics are 'St. Elmo's Fire,' 'Shooting Stars,' 'The lore of Cranburn' (sic) 'petrifying spring, a child that was heard cry n its mother's womb, Elf-stones and Elf-shooting.' Although these miscellanies are of considerable interest, and show Neve to have been a careful observer, less credulous than his neighbours, the inclusion of details in this meteorological paper would be inappropriate.

(*a*) *Terminology*

Neve's vocabulary was quite extensive, and he was very precise in his choice of words, often inserting an alternative in case his first choice was misunderstood. This is well evidenced in the circumlocutory accounts of the aurora borealis :

> 'I likewise observed it often and long, that the sudden clouds of vapours or smoaks . . . they were often varying in their shapes and figures and frequently changeing their places and I could see the fixt Stars thro' the Streamers or Pillars of Light and I took particular Notice of some very bright Rays or Pyramides . . . .'

Some of his words appear only rarely. 'Floody' may be a copyist's error for 'cloudy' or 'flood' or a telescoping of the two. There is little use of dialect. 'Smur' is a local word imported to Ireland from Scotland and 'Fere' is probably another example. The list of words used is given below, classified under suitable headings :

STATE OF SKY

Overcast, overcasting, almost o., almost quite o.
Cloudy, a little c., very c., c. horizon.
Clear (with spellings clair and cleair)
Fair, pretty f., almost f., fairish, f. only ith horizon

CLOUD APPEARANCES

White clouds
White and black clouds
White and curdled clouds
Clouds like smoake
Wav'd and flame-coloured sky
Mackerel sky

WEATHER

Spitting
Droping, a few drops, some drops (or droops)
Misling, misling rain.
Smur, smir, smer (= fine, misty rain)
Rain, rainey, great r., much r., some r., hard r., small r., some small r., great Store of Wind and r
Flood, great f., floody. In a few cases it seems likely that ' flood ' = overflowing of rivers, and elsewhere it signifies merely heavy rain.
Showers, showery, small s., great s.
Hail, some h.
Sleet
Snow, much s., some s., snowing fast, small s., little s.; snowey, very s.
Thunder, thunderstorm, thunder much, 4 (astrologer's symbol for Jupiter)
Lightning, great lightnings
Frost, frosty, hard f., small f., hoar f., hoary f., white f., black f.
Gelu, g. dura, album g. (= Frost, hard f., white f.)
Rime, Rhime, great r.
Thaw, thawey, gentle t.
Dew, great honeydews
Pruina (= dew); P. gelu (= hoar frost)

VISIBILITY

Fog, foggy, fogy, great f., thin f., remote f., f. ith botoms (of valleys), ground f.
Mist, great m., m. ith botoms
Haze, hazey, little h.
Thick air

TEMPERATURE

Hot
Warm, very w.
Cold, very c.
Close
Drying

WIND (was described throughout on a numerical scale, but the following were also used) :

Windy, high wind, great storms of w., some whirlwinds, Stormy

MISCELLANEOUS

Fare and Fere were used a few times. Meaning unknown.

4. Wind

Observations of wind direction and force were included regularly. Direction was normally given on the 8-point compass with only occasional mention of intermediate points, and with no indication of whether the bearings were true or magnetic. Fortunately, this is unimportant, in that the magnetic and geographic meridians were almost coincident over the British Isles at that period.

The diarist gave no details of his scale of wind force and it is probably similar to that given by Defoe (1704) :

(0)	Stark calm	(5)	A topsail gale
(0 or ½)	Calm weather	(6)	Blows fresh
(1)	Little wind	(7)	A hard gale of wind
(2)	A fine breeze	(8)	A fret of wind
(3)	A small gale	(9)	A storm
(4)	A fresh gale	(10)	A tempest

This is the scale which, through Smeaton (1759) was developed into a practical form by Beaufort. No attempt has been made, however, to convert Neve's observations of force into equivalent velocities.

Many of the observations are suspect: there are periods of several days with the same direction and force throughout – probably when the diary was written after the event from a vague recollection that there were no marked changes during the period. On the other hand, when changes were reported, there were a remarkably large proportion of complete reversals of direction.

Tables 1 and 2 summarize the frequencies of direction and include values of resultant direction and its steadiness. The former was calculated assuming all winds to have equal speed, and the latter expresses the ratio of the magnitudes of this vector mean speed and the scalar mean speed. The resultant direction of 200° for the whole year agrees remarkably well with that obtained from modern data (*Climatological Atlas* 1952), the recent value for Aldergrove being 195°.

TABLE 1. Wind directions at Ballyneil 1711-25 (N.S.)

Year	Percentage frequencies									Resultant direction	Steadiness
	N	NE	E	SE	S	SW	W	NW	Calm		
1711	6	6	6	14	20	11	17	12	8	220°	0·21
1712	14	4	5	17	13	16	10	13	8	225°	0·13
1713	19	4	8	18	16	10	7	13	5	180°	0·05
1714	14	0·4	2	14	24	17	15	8	6	215°	0·33
1715	14	2	7	12	28	13	4	14	6	195°	0·21
1716	17	5	6	13	19	11	11	14	4	240°	0·11
1717	12	3	13	5	29	15	10	11	2	205°	0·21
1718	14	4	4	14	30	14	8	8	4	195°	0·28
1719	8	3	7	20	29	16	6	7	4	180°	0·39
1720	10	5	6	9	32	19	5	10	4	200°	0·29
1721	12	8	4	16	21	15	7	13	4	200°	0·17
1722	9	4	6	16	25	16	8	10	6	195°	0·30
1723	9	4	8	13	23	15	11	10	7	200°	0·25
1724	11	4	9	11	26	12	15	5	7	195°	0·26
1725	15	4	14	6	23	9	14	9	6	215°	0·11

TABLE 2. Annual variation of wind directions

Month (N.S.)	Percentage frequencies									Resultant direction	Steadiness
	N	NE	E	SE	S	SW	W	NW	Calm		
January	8	4	5	13	31	17	6	8	8	190°	0·35
February	11	4	7	13	27	19	9	6	4	195°	0·33
March	11	5	8	11	21	14	10	15	5	205°	0·16
April	14	5	12	17	16	14	7	10	5	165°	0·14
May	11	10	12	17	15	11	8	11	5	145°	0·13
June	13	3	6	17	23	15	10	9	4	200°	0·23
July	18	4	4	9	20	10	14	17	4	270°	0·20
August	13	5	8	16	21	13	10	8	6	190°	0·19
September	11	4	6	15	23	12	9	12	8	200°	0·21
October	12	2	6	13	28	14	11	8	6	200°	0·23
November	9	2	5	9	32	15	13	10	5	210°	0·36
December	16	3	5	8	31	11	11	10	5	215°	0·24
Spring	12	6	11	16	17	13	8	12	5	175°	0·13
Summer	15	4	6	14	21	12	11	12	5	215°	0·17
Autumn	11	3	5	12	28	14	11	10	6	205°	0·27
Winter	12	3	5	11	30	16	9	8	6	200°	0·31
Year	12	5	7	13	24	14	10	10	5	200°	0·21

To make the results comparable with the modern data the diary dates were adjusted to 'new-style' calendar (N.S.) by adding eleven days and by beginning the year at 1 January.

Table 3 tabulates the numbers of days on which Neve reported the wind as reaching Force 7 or 8 on his scale. The annual totals range from 3 in 1714 to 15 in 1718 and the monthly totals show that March and December were the windiest months. 15 years' data are not enough to provide a reliable analysis of daily frequencies but they suggest that the dates most subject to gales were 2-4 March (O.S. 19-21 Feb.), 6-9 October (O.S. 25-28 Sept.) and 21-24 December (O.S. 10-13 December). All three agree with Brooks's dates (1946, 1954) for stormy spells in Britain.

A recent analysis of 205 destructive storms near Dublin, spread over several centuries, suggests that the most pronounced singularity of storminess is that at the end of February and beginning of March (Dixon 1959).

TABLE 3. Number of days with winds of force 7 or 8

Year	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1711	0	0	2	1	0	0	0	0	0	1	2	1	7
1712	0	0	3	1	0	0	0	0	0	1	0	0	5
1713	0	1	1	1	0	0	0	0	0	0	1	0	4
1714	0	1	2	0	0	0	0	0	0	0	0	0	3
1715	1	6	2	0	0	1	0	0	0	1	0	0	11
1716	0	0	0	0	0	0	0	0	5	1	1	0	7
1717	4	0	1	1	0	0	0	0	0	0	0	5	11
1718	0	2	5	0	0	1	0	0	0	3	0	4	15
1719	1	1	6	0	0	0	0	0	0	1	0	0	9
1720	3	2	0	0	0	0	0	0	0	1	1	3	10
1721	2	0	0	0	0	0	0	0	0	0	1	1	4
1722	0	0	0	0	0	1	1	0	0	1	3	0	6
1723	0	0	2	0	0	0	0	0	0	0	1	3	6
1724	1	1	1	0	0	0	0	0	1	1	0	4	9
1725	0	1	0	0	0	0	0	0	0	2	0	3	6
Mean	0·8	0·9	1·7	0·3	0·0	0·2	0·1	0·0	0·4	0·8	0·7	1·6	7·4

In Table 4, relating the Force 7 and 8 winds to their directions, the sheltering effect of Slievegallion, to the west of Ballyneil, is apparent. The fewness of occasions with strong winds having an east component is less understandable, and suggests there may have been some local shelter between the site and the open waters of Lough Neagh.

TABLE 4. Frequencies of directions of winds of force 7 or 8

Direction	N	NE	E	SE	S	SW	W	NW
Frequency (per cent)	6	2	3	3	28	27	10	21

5. Precipitation

Neve had a raingauge installed and in operation at the time his diary began in 1711. A covering letter to Dr. Jurin (then Secretary to the Royal Society) includes this description of the gauge :

> ' The method I made use of, to observe the Rain &c that falls, I have a Circular Tin Funnel, exactly twelve Inches Diameter, the which is fixt on the Ridge of my House, remote from any Chimney; to prevent the spattering of Drops (from them) into the Funnel. The nose (or pipe) of this Funnel, goes through the Ridging of the House into a Cock loft (above the Garret) & there I have a Tin pipe about 2 feet long, whose upper end slips on tight, upon the nose of the Funnel, & the lower end as tight into the nose of a large Glass Bottle which receives the Rain, etc. out of the Funnel. When I observe the quantity of Rain which is fallen (at any Time) I have a Cilindrical Glass of about 1½ Inch diameter, & so deep, or long, as to hold something more than a Pound Troy of water : I weighed a Pound Troy of water & put into this Cilindrical Glass & observed the surface of the Pound of water & marked it on the outside of the Cilinder, and divided the space betwixt inside of the Bottom of the Cilinder & the Surface, or Top, of the water, into 100 parts by having a strip of white paper glued on the outside of the Cilinder & divided & numbered from 1 to 100.' .... ' And least I should chance sometimes to omit going to my Bottle fore it be full, I have a Glazed Earthen Pan, that the Bottle stands in : to save any water that may chance to run over the Bottle before I get to it.'

The measurements were thus entered in pounds troy and hundredths. In a summary, however, Neve adopted the modern procedure by converting to equivalent depths, the factor used being apparently 0·200 in. for each pound troy. This is consistent with the dimensions he gives, since one inch in a 12 in. diameter gauge yields 0·06545 cu. ft, which at 50°F weighs 4·963 lb. troy, leading to a conversion factor of 0·2015.

It is obvious that the gauge was over-exposed and that eddy losses were large. Further, it is unlikely that the tin tube was really tight fitting and there were probably losses owing to leaks and evaporation. Evaporation losses were accentuated by the observer's practice of not making a measurement until his bottle was full. After the first two years, however, he developed the habit of measuring the water at the end of each month, even if the bottle was almost empty at the time.

The notebook contains three versions of the precipitation totals. First, there are the actual measurements, entered to the day of measurement and totalled at the end of each month. Second, there is a summary for 1711-24, and third, there is a revised summary for 1711-25, explaining that the other had been copied by ' a Boy, who had made Mistakes.'

I have accepted the annual totals of this final table but have adjusted some monthly figures, partitioning a measurement in the proportion of the numbers of days on which precipitation was reported. Table 5 presents the resultant data, expressed as percentages of the 15-yr averages, since the actual measurements, whether in pounds or inches, are not suitable for comparison with modern data for a gauge having standard exposure. The actual average for Neve's gauge was only 12 in. The site is near the 40-in. isohyet on modern maps.

What can be compared is the annual distribution of the precipitation. Table 5 includes the isomeric coefficients (i.e., monthly values as percentages of the annual) for 1711-1725 and for two recent periods, the latter being derived from climatological atlases (R. Met. Soc. 1926; *Clim. Atlas* 1952). The most striking contrast is that February, one of the driest months in 1711-25, was one of the wettest in 1901-30. The discrepancy may be due to the failure of Neve's gauge to catch snow. All periods agree in showing maxima in December and August.

(*a*) *Rain-days*

Table 6 gives the number of days in each month on which some form of precipitation was reported. Inevitably the totals are smaller (by about one-third) than modern data, since the diarist did not keep constant watch and missed many occurrences, especially at night. It is surprising that he should have recorded as many as 23 rain-days in one month (October 1712).

The uncertainties include the repetition of entries for December-February in 1712 and 1713 (O.S.). In a few cases adjustments have been made, e.g., September 1723 had 9 days with rain reported out of 27 days with diary entries. The 9 has been adjusted to 10.

The distribution of the rain-days differs from that of the precipitation totals, the most obvious disagreements being in the annual totals. An apparent change towards drier conditions in Table 5 is missing from Table 6, suggesting that the trend is spurious, and results from increasing defects in the gauge or in the procedure for measuring.

TABLE 5. ANNUAL DISTRIBUTION OF MONTHLY PRECIPITATION (AS PERCENTAGES OF THE 15-YR AVERAGES)

Year* (O.S.)	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year r
1710/11	23	100	202	98	101	148	40	105	118	174	183	87	113
1711/12	57	34	162	142	84	214	255	100	151	153	127	31	125
1712/13	95	114	77	24	66	104	158	148	41	48	101	95	92
1713/14	59	177	53	49	30	39	274	78	106	184	154	106	114
1714/15	172	82	112	225	278	144	54	97	149	101	42	15	117
1715/16	27	81	46	113	201	88	55	50	286	182	107	60	108
1716/17	143	89	100	126	112	159	88	99	155	20	106	26	99
1717/18	23	128	154	184	125	55	48	354	19	174	97	419	159
1718/19	242	319	59	91	135	56	82	71	63	135	206	110	129
1719/20	220	108	195	33	37	80	69	70	114	130	92	132	103
1720/21	113	17	76	207	60	134	107	111	55	44	92	17	86
1721/22	14	23	127	9	166	139	130	33	24	51	79	152	79
1722/23	92	128	48	47	5	29	69	19	38	60	22	46	48
1723/24	91	55	66	45	37	18	33	92	85	15	31	150	62
1724/25	130	47	22	105	74	92	37	72	95	29	58	52	67
					Monthly as percentages of annual averages								
1711-25	7·7	6·3	6·2	7·9	8·2	5·6	9·8	11·2	8·4	7·9	11·1	9·8	
Adjusted to N.S.	8·4	6·9	6·2	7·3	8·1	6·6	8·3	10·7	9·4	8·1	9·9	10·3	
1881-1915	8·2	7·4	7·8	6·6	7·2	7·2	8·9	10·6	7·6	8·8	9·0	10·7	
1901-30	8·6	9·8	7·1	6·8	7·4	7·1	8·1	10·1	6·9	9·7	8·6	9·8	

TABLE 6. NUMBERS OF DAYS WITH REPORTED PRECIPITATION

Year* (O.S.)	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1710/11	5	16	18	14	9	11	7	14	12	12	12	16	146
1711/12	10	11	12	10	9	9	21	12	16	23	22	(7)	162
1712/13	(14)	(8)	10	10	13	7	18	15	6	10	16	(7)	134
1713/14	(14)	(8)	8	12	11	5	14	17	8	21	9	10	137
1714/15	10	13	11	19	24	15	13	16	17	11	7	6	162
1715/16	9	10	15	10	12	7	8	7	20	14	12	13	137
1716/17	7	9	16	11	11	7	11	11	16	5	7	13	124
1717/18	5	9	11	12	6	8	4	19	5	11	8	16	114
1718/19	9	10	10	7	7	8	18	10	12	14	11	9	124
1719/20	15	11	12	3	10	13	13	20	16	14	17	12	156
1720/21	15	5	13	18	13	10	13	15	9	7	20	4	142
1721/22	16	10	13	8	18	14	16	14	8	9	17	11	154
1722/23	8	19	9	9	5	8	12	7	10	6	13	14	120
1723/24	14	12	14	13	13	11	15	17	12	11	13	22	167
1724/25	15	5	8	17	17	17	11	14	14	12	9	12	151
Mean	11·1	10·0	12·0	11·5	11·8	10·0	12·9	13·9	12·1	12·0	12·9	11·5	142
Mean adjusted to N.S.	11·2	10·4	11·6	11·7	11·7	10·7	11·9	13·5	12·8	12·0	12·6	12·0	142
per cent	7·9	7·3	8·1	8·2	8·2	7·5	8·4	9·5	9·0	8·5	8·9	8·5	100
1901-30 per cent	9·2	7·9	8·3	7·5	7·9	6·6	7·9	9·2	7·5	9·2	9·2	9·6	100

(The lines 'adjusted to N.S.' in Tables 5 and 6 were obtained by assuming uniform distribution through each month, e.g., adjusted September value $= \frac{11\,A + 19\,S}{30}$ where A, S are the values for the O.S. August and September).

* Before the change from O.S. to N.S. in 1752 the calendar year ended on 24 March. Dates between 1 January and 24 March O.S. would therefore fall in the previous year as compared with N.S. reckoning. In Tables 5 and 6 therefore, each row covers parts of two calendar years (O.S.), denoted thus : 1710/11.

An isomeric comparison is again possible and the major difference is an apparent ncrease in the frequency of precipitation for each month from October to January nclusive. This can be explained by the longer time Neve spent in bed during the dark inter months, and allowing for this human factor there is no evidence of any significant nange in the 200 years.

6. Snow

A few of Neve's snow reports are ambiguous, e.g., ' Snow 1 n ' at the morning observaon, may report the depth of fresh snow remaining from a previous fall and not necessarily resh snow. Such reports have been accepted as ' days with snow ' for this summary, vhich omits, however, the mentions of snow on the mountains when none was noted at ow levels. No attempt has been made to compute the numbers of days with snow lying.

TABLE 7. Number of days with snow (N.S.)

ear	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
711	(0)	7	7	0	0	0	0	0	0	0	0	1	15
712	4	5	3	0	0	0	0	0	0	0	4	(1)	17
713	(2)	(5)	0	3	0	0	0	0	0	1	4	(1)	16
714	(2)	(5)	0	4	0	0	0	0	0	0	0	2	13
715	0	0	3	0	1	0	0	0	1	0	0	2	7
716	6	0	4	0	0	0	0	0	0	0	0	4	14
717	3	1	7	0	0	0	0	0	0	0	0	4	15
718	6	2	4	4	0	0	0	0	0	0	0	2	18
719	0	1	0	0	3	0	0	0	0	0	0	1	5
720	3	3	0	0	0	0	0	0	0	0	1	4	11
721	3	8	3	0	0	0	0	0	0	0	4	1	19
722	0	4	1	1	0	0	0	0	0	0	3	7	16
723	0	1	1	0	0	0	0	0	0	0	0	1	3
724	1	5	4	0	1	0	0	0	0	0	1	3	15
725	6	0	0	1	0	0	0	0	0	0	1	4	12
otals	36	47	37	13	5	0	0	0	1	1	18	38	196
verage 711-25	2·4	3·1	2·5	0·9	0·3	0·0	0·0	0·0	0·1	0·1	1·2	2·5	13
verage 912-38	3·7	3·2	3·6	1·4	←		2·4			→	1·0	2·7	·17

Allowing for the difference in the frequencies of observations, the average of 13 days er year is consistent with the modern estimate of 17 days per year. The deficiency is most narked in January and March. Ballyneil had no snow from June to August and only once ach in September and October. The snowiest winter was 1717-18, and that most free from now 1718-19. The heaviest snowfall was at the beginning of 1721 (N.S.) when it reached depth of one foot and was still lying in ditches at the end of February, and on mountains ntil the end of March. ' There was a great many Cattle lost in this Storm on the Mounains and for want of Fodder.'

7. Frost

Table 8 summarizes the numbers of days with mention of frost at one or more of the bserving hours. The range is rather large, from only 17 days in the mild year 1722, o 54 days in the immediately preceding year.

The frostiest winter was 1715-16 when more than half the days from December to ebruary experienced frost. In a note at the end of the February diary Neve wrote :

> ' NB. This frost was so hard that Lough Neagh was frozen so hard yt (as Robert Simpson told me) his son & some other boys went on it a gt. way from the shore : & at Colrain the River Band was frozen so hard that People walked over it : & I was told that People walked over Lough Beg. NB. There hath been & is a great death of sheep since ye wet summer.'

TABLE 8. Number of days with frost (N.S.)

Year	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total
1711	(6)	16	2	2	0	0	0	0	0	1	6	6	(39)
1712	11	7	9	1	0	0	0	0	0	3	9	(12)	(52)
1713	(2)	(8)	1	1	0	0	0	0	0	1	11	(14)	(38)
1714	(2)	(8)	2	1	3	0	0	0	0	0	1	9	(26)
1715	7	3	4	2	0	0	0	1	0	3	4	10	34
1716	18	15	3	1	0	0	0	1	1	0	0	10	49
1717	14	8	7	0	0	0	0	0	0	0	3	5	37
1718	14	5	0	5	0	0	0	0	0	4	2	10	40
1719	14	3	4	3	4	1	0	0	0	0	5	1	35
1720	4	4	3	1	0	0	0	0	0	2	3	8	25
1721	2	19	18	4	0	0	0	0	1	0	7	3	54
1722	3	2	3	0	0	0	0	0	0	0	2	7	17
1723	8	3	1	1	0	0	0	0	0	2	2	10	27
1724	5	14	11	0	0	0	0	0	0	4	3	7	44
1725	10	2	5	3	0	0	0	0	0	1	10	6	37
Average	8·0	7·8	4·8	1·7	0·5	0·1	0·0	0·1	0·1	1·4	4·5	7·9	36·9

No comparison with modern data has been attempted.

8. Pressure

Barometer readings began in June 1722 and were made thrice daily thereafter, with occasional short breaks, the longest interruption lasting nine days in May 1723. The barometer, constructed by Johannes Cassina, was of the J-tube design. As its owner wrote:

> 'This Glass hath no Sisteren but its curved or bent at the lower end where is it were a vial its devided from the 28th inches to the 31 inches into 36 parts or Devisions and against the number stand in the Marge.'
>
> 36 great Drought
> 30 settled Fair
> 24 Fair Weather
> 18 Changeable Weather
> 12 Rain or Wind
> 6 Very Rainey
> 2 Storm
>
> .... the tubes about ¼ of an Inch Diameter at the outside and as I guess about ⅛ in. within.'

As with other elements, the observations were not impeccable: possibly because of friction in the narrow tube, the day-to-day changes are suspiciously small, e.g., the readings were continuously '23' from mid-day on 21st to mid-day on 24th November 1725. The accuracy of reading as usually to the nearest division (= 1/12 in.) sometimes to the half division, occasionally to the quarter, and for one short period to the eighth.

The summary in Table 9 was obtained by averaging all the observations in each month (after rewriting with a New Style calendar) and converting the result to millibars using 30 in. = 1015·917 mb.

TABLE 9. Mean atmospheric pressure in millibars (N.S.)

Year	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Annual mean
1722						995	989	994	999	1006	992	994	(996)
1723	1006	996	992	1001	1010	1004	1010	999	1008	1011	999	994	1002
1724	991	991	997	1004	1002	1004	1001	1001	1004	1006	1008	991	1000
1725	993	1017	1016	1001	1000	997	998	997	1007	998	1002	1000	1002
Mean	996	1001	1001	1002	1004	1000	1000	998	1005	1005	1000	995	1001
Max.	1020	1023	1022	1016	1024	1016	1019	1016	1019	1022	1019	1019	1024
Min.	954	972	976	982	973	974	968	981	974	971	957	962	954

Close inspection of the data does not reveal any regular diurnal or annual fluctuation consistent with there being a serious temperature coefficient.

The modern 'normal' of mean sea-level pressure in the area is about 1012 mb, about 11 mb higher than the annual mean deduced for Ballyneil 1722-25. About 8 mb of this difference can be ascribed to the effect of the height above m.s.l., the site being near the 200 ft contour, but it is not possible to say how much of the residual 3 mb is due to climatic change and how much to instrumental error.

9. Temperature

The thermometer was made by Cassina from glass tubing similar to that of the barometer, and contained 'spirits of wine.' It was graduated from − 100° to + 75°, the zero being Temperate, + 60° – Very Hot, + 40° – Hot, + 20° – Warm, − 20° – Cold, − 40° – Colder, − 60° – Very Cold and − 80° – Excessive Cold. The readings show so large a non-cyclic change (about $\frac{1}{2}$° per month) that it is clear the spirits were unstable, and it is not possible to decide what fixed points the maker used for his graduations.

The location of the thermometer is not stated, except that it was near the barometer. They were presumably indoors in an unheated room. The average diurnal range between morning and midday observations was only 5° and fluctuated erratically. To add to the difficulties of interpretation the diary includes many obvious errors of copying and some uncertain signs. Neve at first used ⌜ after a number to denote positive and ⌝ before it for negative, but they were sometimes reversed, or replaced by : or omitted. Such slips were so numerous in the first few months of the record that the data for May to December 1722 have been rejected, and the summary is confined to the three complete years 1723 to 1725.

The procedure adopted in preparing the summary was to evaluate a mean temperature for each day as the arithmetic mean of the three observations, interpolating for missing observations with the aid of the diurnal trend shown by neighbouring days. The resultant daily means were then averaged for each month (New Style calendar). A non-cyclic correction was applied, 0·5° for each month since the start of the record. Finally the adjusted monthly means were converted to the Fahrenheit scale using the empirical formula $F = 49 + \frac{5}{12} N$, according to which the limits of the graduations would correspond to 80 and 7°F respectively. The final results are shown in Table 10, which includes extremes as well as means.

TABLE 10. Mean and extreme temperatures in °F (N.S.)

Year	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Annual
						Mean							
1723	36	41	42	44	52	61	59	62	59	53	48	43	50
1724	42	37	40	46	53	60	59	61	58	47	43	40	49
1725	39	43	44	48	52	55	58	57	60	49	46	42	49
						Extreme maximum							
1723	46	47	47	57	67	68	63	68	65	61	55	53	68
1724	51	45	47	54	66	69	66	68	66	52	53	52	69
1725	47	47	52	56	62	62	65	62	67	59	55	50	67
						Extreme minimum							
1723	28	29	31	37	33	54	54	57	49	41	41	33	28
1724	34	31	33	38	44	51	50	54	43	38	32	31	31
1725	33	36	36	33	44	48	48	52	50	39	33	33	33

As the procedure assumed that the annual mean differed little from that prevailing in the same area today, the final figures have few applications. It is noteworthy, however, that the fluctuations are in good agreement with Manley's data for central England – especially the cold January 1723, February and October to December 1724, and August 1725. (Manley 1953).

(*a*) *Cold singularities*

It is widely accepted that there are significant dates when the temperature tends to be regularly above or below the seasonal average, although the dates are not sufficiently fixed to be of great value to forecasters. Manley (1958) has listed seven cold periods detectable in his analysis of snowfall reports 18-23 November, 15-16 December, 7-11 January, 1-4 February, 19-20 February, 9-12 March and 11-14 April. An independent estimate from the Ballyneil data of 1711-1725 yielded as the dates for cold spells 14-18 December, 30 December-1 January, 5-8 February, 18-20 February, 2-4 March, 11-13 March and 13-15 April. The frequency of frost reports at Ballyneil suggests 12-17 December, 31 December-2 January, 12-15 January and 18-23 February. Thus there is good support for Manley's December period and the second February period, but the evidence for others is unsatisfactory. Of the Ballyneil spells only that of late February is in Brooks's list (1946, 1954).

10. Thunder

Neve mentioned thunder on 69 days and lightning on many others. The most thundery years were 1722 and 1723, each with 8 days, and at the other extreme 1714 had only one day. The average of 4·6 days per year is in excellent agreement with 5·1 for Armagh 1881-1915 or with 6·2 deduced from maps for 1901-30 (*Clim. Atlas* 1952). The annual variation also shows no significant change in the 200 years : two-thirds of the thunderstorms occurred in the months June to August, one-sixth in March to May and the remainder were spread over Autumn and Winter.

Below are given some of the supplementary notes to the diary which include details of damage done during thunderstorms. The dates here are the original (O.S.) dates.

> 1717 22 July. There were two men killed Dead at Dunganon & a Deale of Cattle with the Thunder
>
> 1718 3 August. A mountain broke above Garavagh and the water carried away a House.
>
> 1721 6 June. The Thunder split a Tree in John Orr's land & set it on fire.
>
> 1723 22 May. This p.m. towards Night was some Thunder (& rain at some places); at Ballygun it Rained a grt shower about ½ a Mile of us and here we had only a few drops, that wd not have Wett thro' a Handkerchief but it is to be observed wr it did it was nearer to the Mounts. where the Thunder was that day.

TABLE 11. Frequency of thunder (N.S.)

Year	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total
1711	0	0	0	0	0	1	0	0	1	0	1	1	4
1712	0	0	0	0	0	1	0	1	0	0	1	0	3
1713	0	0	1	0	1	2	0	0	0	0	0	0	4
1714	0	0	0	0	0	1	0	0	0	0	0	0	1
1715	1	0	1	0	0	1	0	0	2	0	0	0	5
1716	0	0	0	0	0	1	1	0	0	0	0	0	2
1717	0	0	1	0	1	0	2	3	0	0	0	0	7
1718	0	0	0	0	0	2	0	4	0	1	0	0	7
1719	0	0	0	0	0	1	1	2	0	0	0	0	4
1720	0	0	0	0	0	1	1	2	0	0	0	1	5
1721	0	1	0	1	0	1	0	0	0	0	0	0	3
1722	0	0	0	2	0	4	2	0	0	0	0	0	8
1723	0	0	0	1	1	1	1	4	0	0	0	0	8
1724	0	0	0	1	1	0	0	1	0	0	0	2	5
1725	0	0	0	0	0	1	0	0	2	0	0	0	3
Average	0·1	0·1	0·2	0·3	0·3	1·2	0·5	1·1	0·3	0·1	0·1	0·3	4·6

	Spring	Summer	Autumn	Winter	Year
Ballyneil 1711-25	0·8	2·9	0·5	0·4	4·6
Armagh 1881-1915	1·0	3·7	0·2	0·2	5·1
Ballyneil 1901-30	1·0	4·4	0·5	0·3	6·2

11. Hail

Statistics of the frequency of hail are of dubious value owing to the uncertainty of distinguishing between hail and snow, and the difficulty of recognizing a few hailstones intermingled with heavy rain. Even taking these points into account, and considering that Neve's observations were almost entirely confined to the daylight hours, his average of only 2·3 days per year with hail seems rather small by comparison with modern observations.

Another noteworthy feature is that of Neve's 35 mentions of hail, only 3 were during the summer, the season which now has the maximum frequency. The maximum for 1711-25 was a well-marked one in April-May.

TABLE 12. Number of days with hail (N.S.)

Year	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1711	0	0	0	0	0	0	0	0	0	0	0	0	0
1712	0	0	0	0	1	0	0	1	0	0	0	0	2
1713	1	0	0	1	1	0	0	0	0	0	0	0	3
1714	1	0	0	1	1	0	0	0	0	0	0	0	3
1715	0	0	0	0	1	0	0	0	1	0	0	0	2
1716	0	0	1	1	0	0	0	0	0	0	0	0	2
1717	0	0	0	0	1	0	0	0	0	0	0	0	1
1718	0	0	0	2	0	0	0	0	0	1	0	0	3
1719	0	0	1	2	0	0	0	0	0	0	2	0	5
1720	0	0	0	0	0	1	0	0	0	0	1	0	2
1721	0	0	0	2	2	1	0	0	0	0	0	1	6
1722	0	1	0	0	0	0	0	0	0	0	1	0	2
1723	0	0	0	0	0	0	0	0	0	0	0	0	0
1724	0	0	2	0	1	0	0	0	0	0	0	0	3
1725	0	0	0	1	0	0	0	0	0	0	0	0	1
Average	0·1	0·1	0·3	0·7	0·5	0·1	0·0	0·1	0·1	0·1	0·3	0·1	2·3

	Winter	Spring	Summer	Autumn	Year
Ballyneil 1711-25	0·3	1·5	0·2	0·4	2·3
Armagh 1881-1915	0·2	1·4	3·7	0·2	5·1

12. Aurora borealis

The period of Neve's diary was a time when the sun had again become active after its seventeenth-century quiescence, and the associated displays of the aurora borealis included some spectacular occurrences. Neve had described the notable display of 16 November 1707 (Boate and Molyneux 1726), and gives details of several of the 15 observations during the period 1711-1725. Many are dates additional to Angot's list (Angot 1896).

In 1716 ' strange light in the North ' was reported on 31 March and there are similar entries for the succeeding two nights. ' Strange light ' on 2 November of the same year is the only other report before 1719 when, on 22 December, ' about 9 and 10 it was very Light above the clouds ith NW. Horizon.'

A display in the early hours of 10 November 1720 was so bright that some people left their beds thinking that morning had dawned. On ' 6 February 1720 a brilliant rayed arc and corona was visible as early at 7.45 p.m. but were short-lived, there being only a faint arc left soon after 8 p.m.' On 11 September 1721 ' Many Northern streamers appeared ' and on 7 February 1721/22* the entry was ' N. streamers.' The entry for 4 October 1722 is much more informative :

* See footnote on p. 378.

'This morning I got up about 2 a Clock (tho' I went to Bed late) . . . it was so light that I thought it was crepusculum of the Morning;' Almost all types of aurora seem to have developed with very rapid transformations, and 'it aprd, as we was standing at some distance from a Batalion of Men viz opposite to one of ye Flanks, and some of the files seemd to be marching very swiftly toward ye N and other files contrary marching by them as swiftly toward the West, or like people going to or from a fair or Market and are Runing to it, some runing from it, all as if they were in great heat (or as wife compared it to a countrey Dance; for I called up all my family to see it, for it was something Terrifieing to behold, for 1/3 of the Heavens seemed sometimes to be a Blaze of fire – there was twelve of us looking at it, & all seemed to be in a kind of consternation : one of my Servants that was a Irish papish : yt was so affryted yt he was going to his prayers for he thought the day of Judgement was come.'

The next few reports are shorter :

9 October 1722. 'Some streamers in the NW.'

5 November 1722. 'This morning Mr. Christ. Hanbury told me yr was very strong Light flashing in the sky N & NW which continued near 4 hours . . . .'

21 February 1722/23*. 'This night there was some Streamers.'

22 March 1722/23*. 'N. Streamers.'

26 September 1725. 'This night betwixt 8 and 9 a Clock I observed an unusual Appearance of ye Aurora Borealis . . . the bright Rays of Light seemed to proceed or Stream from two Black Clouds one in the East & ye other in ye western Parts of the Horizon . . . they passed quite thro ye zenith & Appear'd like 3 & sometimes 4 Semicircles of Light wch had a Vibrating or Undulating Motion . . .

(*a*) *27-day period*

It is widely known that the solar rotation period of about 27 days leads to a recurrence interval of the length in the dates of auroral displays. The periodicity can persist long after associated solar features have died out, and there is some evidence that it may continue unchanged for many years (Dixon 1939). Neve's data fit very well to the sequence 1719 21 December + 27·606 N days. Of the 10 occurrences in the years 1719 to 1723 seven fit the sequence within 2 days and none as much as a week out.

Actual date		N	Theoretical date	Difference
1719	Dec. 22	0	Dec. 21	− 1
1720	Nov. 18	12	Nov. 16	− 2
1720/21*	Feb. 6	15	Feb. 7	+ 1
1721	Sept. 11	23	Sept. 16	+ 5
1721/22*	Feb. 7	28	Feb. 1	− 6
1722	Oct. 2	37	Oct. 7	+ 5
1722	Oct. 9	37	Oct. 7	− 2
1722	Nov. 4	38	Nov. 4	0
1722/23*	Feb. 21	42	Feb. 22	+ 1
1722/23*	Mar. 22	43	Mar. 22	0

The 27·606 is rather longer than one usually finds. Twentieth century data suggest about 27·3 days and the dates given by Angot (1896) for the 1716-23 auroras lead to 1716 20 January + 27·34 N, which does not fit to Neve's dates.

13. Miscellaneous

Optical phenomena

'Iris' — rainbow is mentioned a few times in the diary, and 'Halo' or 'Hallo' appears occasionally, but the appearance was more often recorded by a symbol – a dotted ring around the device for the sun or moon. The only sky colouration mentioned is redness of the sky at sunrise or sunset.

* See footnote on p. 378.

St. Elmo's Fire

On 10 October 1710 ' One Francis Serson and I being comeing homeward in the Night in a townlamd called Carraloon (betwixt Macherafelt and Loughneagh), it grew so extream Dark upon us that for near a Mile's Riding we could see nothing but our Horses Ears, the tips of which seemed to be all on fire : but there came on a great Storm.'

Seiche

On 30 July 1722, ' I observed was a Very Rainey A.M. and the night before to witt⊙ Night the Lough Rored or made a grt Noise tho' but Little Wind sturing.' This is of some historical interest in that Lough Neagh has been the site of modern Seiche studies (Darbyshire 1957).

Exceptional growth

Neve found it noteworthy that in October 1723 he had a second crop of white roses and ' Capt. McGowan told me the 28 day they had a second crop of Apples; they were as big as the top of his finger.'

In the following June ' Spring was very remarkable for fruit Trees Blossoming nay even all kinds for both the black and white Thorns were Clothed with Blossoms for our quicksett hedges were all as white as a sheet the like was never known within Memory of any Man living.' The profusion of blossom led to profusion of fruit and in October 1724 Neve wrote ' Ys year yr was abundance of Apples, Pears, &c. and the Whitethorn Hedges were all Red with Haws.'

Acknowledgments

Permission to publish this paper has been accorded by the Director of the Irish Meteorological Service and the Secretary of the Department for Industry and Commerce. The author wishes to thank the Royal Society for permission to print the extracts from the Neve manuscript weather diary in their possession.

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551.521.14

The reflection of short-wave radiation by vegetation

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(Manuscript received 15 January 1959; in revised form 7 May 1959)

SUMMARY

The fraction of short-wave radiation reflected from agricultural crops was measured from May till September 1958 using two portable solarimetric thermopiles. Maximum reflection coefficients for grass, lucerne, potatoes, sugar beet, and spring wheat were between 0·25 and 0·27. Lower values found in the early stages of crop development, and for spring wheat even at maturity, can be correlated with low leaf areas and mutual shading of the leaves, also reduced reflection. Data by Billings and Morris (1951) have been used to compute total reflection coefficients for different types of vegetation including two desert species and to explain why these do not normally differ as much as the coefficients for the visible component.

1. INTRODUCTION

In a recent study of the annual heat balance of a short-grass surface, Roach (1955) reports that the mean incoming radiation at Kew from May 1953 to April 1954 was 10·4 mw cm^{-2} but the net radiation available for evaporation and for heating the environment was only 4·0 mw cm^{-2}, since 3·8 mw cm^{-2} were lost by long-wave exchange with the atmosphere and a further 2·6 mw cm^{-2} by reflection of short-wave radiation. These figures emphasize the importance of an accurate reflection coefficient in assessing the heat balance of a vegetative surface. For example, to apply Roach's data to other types of vegetation it is necessary to know the reflection coefficient α within $\pm$ 0·02 to calculate net radiation to $\pm$ 5 per cent, given short-wave radiative income and long-wave loss.

Appropriate reflection coefficients cannot always be obtained from the literature and wide variations are reported by different workers for the same crop, depending on density of cover and on weather. For example, the following ranges have been reported for short grass : 0·22 to 0·33 Ångström (1925); 0·14 to 0·37, Kalitin (1930); 0·22 to 0·50 Blackwell quoted by Roach (1955).

Over the last thirty years many studies of reflection coefficient have been made in the U.S.S.R., for which reference can be made to Fritz and Rigby (1957) and to Budyko (1958) who give the following range of values for agricultural crops : rye and wheat 0·10-0·25; potatoes 0·15-0·25; meadows 0·15-0·25; cotton 0·20-0·25. Thornthwaite (1957) reports that the reflection coefficient of common vegetables is very similar to that of grass, e.g., spinach, $\alpha = 0{\cdot}26$. That all these crops should have a similar upper limit suggests that individual leaves of different species may have approximately the same coefficient, achieved for the crop as a whole when there are sufficient leaves to shade the ground completely. This is supported by observations on the detached leaves of ten wild and cultivated species reported by Kondrat'ev (1954) where α lay between 0·22 and 0·27. Equality of reflection coefficient for different crops is implicit in the concept of potential transpiration (Penman 1956); that all short, green crops completely covering the ground receive the same net radiation, and transpire equally in the same weather conditions when soil water is adequate.

To check the hypothesis of a maximum α independent of species for fully-developed crops and to study the way in which α changes with season during crop development reflection from grass, wheat, sugar beet, potatoes and lucerne were measured throughout the growing season in 1958.

2. Apparatus

Commercial solarimeters designed for the routine observation of incoming short-wave radiation at a fixed point are unsuitable for field work because of their weight, which makes them difficult to support above a crop with a readily portable mast. The instrument designed for the present work could be mounted on a light tripod, and output was sufficient to operate a portable millivoltmeter (Monteith 1959).

Before attempting to determine α, two similar instruments were set horizontally with a spirit-level, both facing upwards, to obtain their relative sensitivity. One instrument was then turned to face the ground, receiving radiation reflected from the crop, and the ratio of signals corrected by the relative sensitivity factor then gave α. Five to ten readings were taken on each instrument depending on the constancy of radiation and the standard error of the mean ratio was generally of the order of 2 or 3 per cent. In bright sunshine the reading error on the meter scale (which was divided into 100 parts) was about $\alpha/50$, while in overcast skies it was sometimes as high as $\alpha/15$. The reliability of the determinations of α is therefore a function of radiation intensity. Differences in α of less than 0·01 may not be real and will be ignored in discussion of results. Equipment for calibration in an inverted position was not available but close agreement with the results of other workers who used standard equipment to measure reflection from short, green grass gives confidence in the figures for other surfaces.

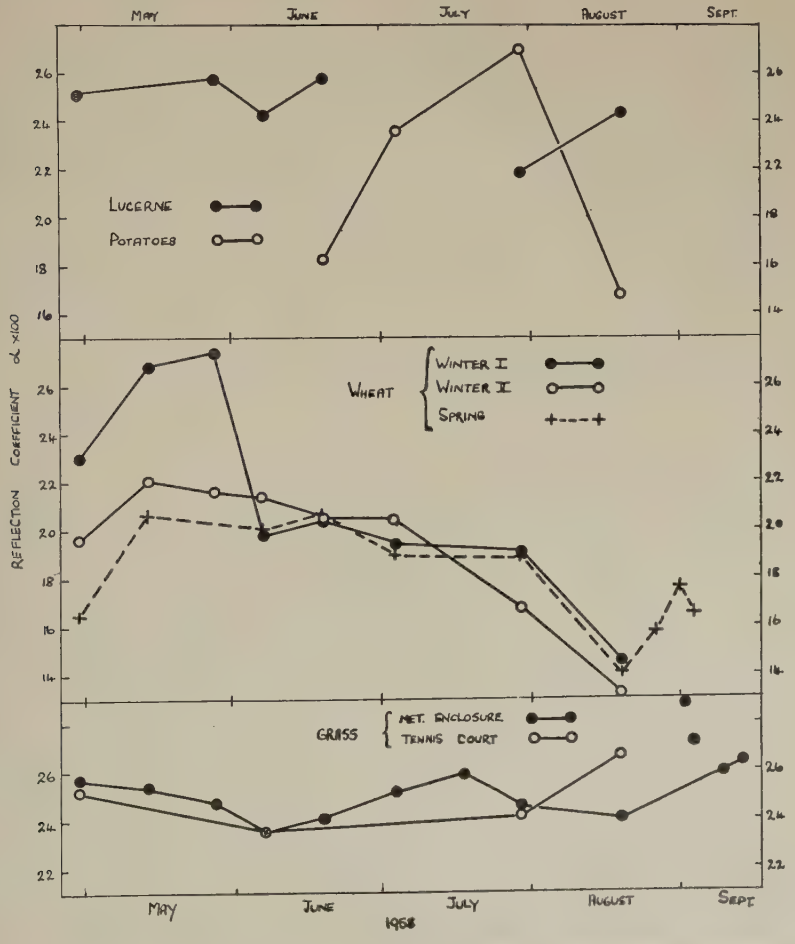

Figure 1. Seasonal variation in reflection coefficient of grass, wheat, potatoes, and lucerne in 1958 (see text).

3. The reflection coefficient of crops

(*a*) *Bare soil*

During the early stages of growth, reflection was mainly from the bare soil between plants, and measurements were therefore made over uncropped ground to determine the range of reflection coefficients that could be obtained with different soil moistures. The soil is a clay loam with flints, and at field capacity with a moisture content of about 35 per cent dry weight, α was 0·11. Measurements after several days of rapid drying without rain gave $\alpha = 0{\cdot}18$.

(*b*) *Grass*

The seasonal variation in reflection coefficient of grass and other types of vegetation is shown in Fig. 1. With the exception of two mornings of heavy dew (1 and 3 September) the reflection coefficient of the grass in the meteorological enclosure lay between 0·24 and 0·26 with a mean of $0{\cdot}25_2$, whereas Roach gives the mean for one month as $0{\cdot}25 \pm 0{\cdot}01$. There was remarkable constancy throughout the season, which was so wet that there was little change in grass colour. For dewed grass the highest observed value was 0·29 with clear sky and solar altitude 35°. The grass in the enclosure, cut approximately weekly to a length about 1 cm, contains a high proportion of other species, e.g., clover, but a lawn tennis court with a closely cut stand of pure grass reflected the same amount of radiation, the average coefficient from four observations between May and August being $0{\cdot}24_9$.

(*c*) *Wheat*

From the end of April until the beginning of June there were differences between the reflection coefficients of two plots of winter wheat and one of spring wheat. The spring wheat was sown on 25 March and its coefficient increased from 0·16 at the end of April, a value characteristic of the soil rather than of the crop, to 0·20-0·21 in May and June. When the crop ripened at the beginning of August there was a sharp drop in reflection to $\alpha = 0{\cdot}14$ followed by a slight increase attributable to vigorous weed growth which had completely covered the ground by the beginning of September.

Winter wheat I was sown on 11 November 1957 and received ammonium sulphate (3 cwt/acre) on 17 March. A very high reflection coefficient in May, about 0·27, dropped suddenly to 0·20 at the beginning of June, and this value was maintained until ripening in August. Winter wheat II, sown on 6 December 1957 received no nitrogenous fertilizer and was only 11 cm tall on 5 May compared with 19 cm for winter wheat I. The highest value of a α for this plot was 0·22, and there was a decrease with ripening before the end of July.

Differences in reflection coefficient between different stands of the same crop may be produced by differences in leaf area, conveniently specified as LAI. Leaf-area index is the ratio of leaf area (i.e., half the total leaf surface) to the area of underlying ground. Measurements by Watson (1947) and by Thorne and Watson (1955) suggest that the LAI of autumn-sown cereals may be significantly greater than for spring-sown cereals and that it can be further enhanced by an application of nitrogen in early spring. In one experiment with winter wheat which had received nitrogen in late April, LAI reached a maximum of 4·5 in the last week of May, but this fell to 3·0 a week later. A similar sharp decrease in leaf area may have been responsible for the decrease in α for winter wheat I at exactly the same time of year.

Differences in available nitrogen also produce marked differences in crop colour, and on 6 June winter wheat I was a much darker shade of green than winter wheat II. On this date, however, there was no significant difference in α suggesting that the reflection of short-wave radiation is not closely linked to reflection of the visible component.

Similar differences in the reflection coefficient of spring and winter sown wheat have been observed by Galperin (1949) and by Berezina (1957) who also observed a diurnal variation, e.g. with winter wheat, from 0·20 at noon with a solar elevation of 64° to 0·32 at 14°.

(*d*) *Lucerne*

Reflection from a mature stand of lucerne sown in the spring of 1957 and growing to a height of about 50 cm was measured on four dates from April until June 1958, and the mean value of α was $0{\cdot}25_2$. A cut was taken on 18 June, but at the end of July there was sufficient re-growth to give $\alpha = 0{\cdot}22$ and on 19 August, $\alpha = 0{\cdot}24$. As with grass, foliage of the mature crop is sufficiently dense to give complete ground cover.

(*e*) *Sugar Beet*

Sugar beet (variety, *Klein E*) was drilled on 11 April and germinated a fortnight later. Until the beginning of July, reflection was mainly from the soil and α decreased when a wet June succeeded a dry May. In the middle of June, after singling, the plants were about 23 cm apart in rows 60 cm apart, and about half the ground was shaded by leaves. By mid-July the plants were almost meeting between the rows and subsequently there was complete cover, with crop height 50-60 cm.

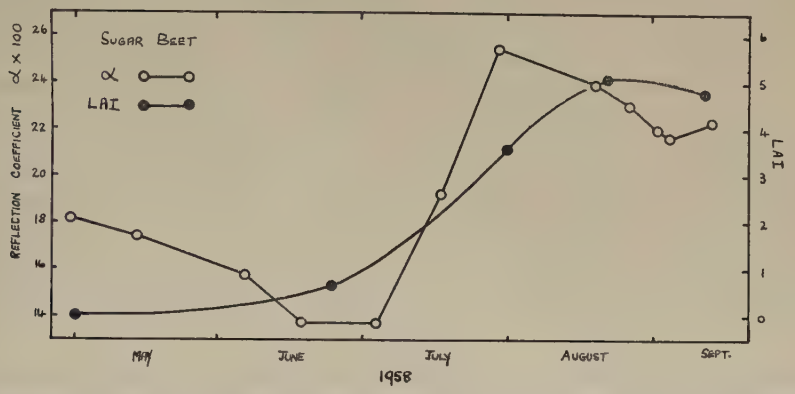

Figure 2. Seasonal variation in reflection coefficient and in leaf-area index (LAI) for sugar beet in 1958.

A more precise description of growth can be given in terms of LAI which was measured for a separate experiment, and observations are plotted in Fig. 2. There is a striking increase of LAI after the beginning of July, matched by an increase of α to a maximum of $0{\cdot}25_5$. Comparison of the two curves suggests that reflection did not increase significantly until LAI exceeded 1·5. This may be because the young leaves of a sugar beet plant grow upwards rather than outwards and at LAI = 1 there is comparatively little shading of the ground. It is only when the older leaves of the plant begin to fall into a more horizontal position that reflection shows a marked increase over the bare soil value. The general vertical habit of sugar beet leaves, which is encouraged when the plants are closely spaced, may also explain the decrease of α after the end of July at a time when there was little change in LAI. Observations suggest that the leaves growing upward from the centre of the plant shaded both each other and more horizontal leaves. Effectively, there was a trapping of radiation between the leaves and it is reasonable to suppose the reflection from such a system is less than for a crop such as potatoes whose leaves have a more dominantly horizontal habit. Since a lower sun would cast longer shadows, α observed at approximately the same time each morning would decrease after midsummer.

The true reflection coefficient of the leaf material was determined by stripping a number of the larger leaves and laying then flat on a short grass surface over an area of about 1 m^2. The radiation reflected from this area (corrected for reflection from the grass after transmission through the leaves) gave $\alpha = 0{\cdot}26$, supporting the hypothesis that lower crop values of α are a consequence of leaf arrangement rather than of leaf colour.

(*f*) *Potatoes*

Potatoes (variety *Majestic*) planted on 21 April in drills spaced at 70 cm gave $\alpha = 0{\cdot}18$ on 18 June when the foliage covered roughly half the ground between rows. By 29 July α had reached a maximum of 0·27 but this fell rapidly to 0·17 on 19 August after the rapid spread of potato blight and collapse of the crop canopy.

With the exception of spring wheat and late-sown unfertilized winter wheat, all crops achieved a similar maximum value of α at or about the time of maximum leaf area.

Grass	$0{\cdot}25_2$ (mean)
Lucerne	$0{\cdot}25_2$ (mean)
Winter wheat (II)	$0{\cdot}27_4$ (27 May)
Sugar Beet	$0{\cdot}25_5$ (29 July)
Potatoes	0·27 (29 July)
Mean of five crops	0·26

The scatter of these values is of the same order as the error inherent in the observations and may not be significant. It appears that, provided there is sufficient leaf area and provided leaves are so arranged that they completely shade the ground without shading each other, crop reflection coefficients are close to 0·26. In practice, two factors can give smaller values of α : (*i*) leaf area inadequate to shade the ground (e.g., spring wheat) so that some reflection takes place from the soil. Observed reflection should then depend on LAI, on the value of α for the soil and on time of day, increase of α being anticipated with decreasing solar altitude, since less radiation penetrates to the ground the lower the sun (Berezina 1957) ; (*ii*) vertical rather than horizontal distribution of plant material, so that there is mutual shading between different parts of the plant (e.g., sugar beet). With decreasing solar altitude, α would be expected to decrease. This effect is probably most marked in forests which are known to have lower values of α than agricultural crops, e.g., coniferous forest 0·10-0·15; deciduous 0·15-0·20 (Budyko 1958). The particularly low values for conifers are often related to the dark colour of the foliage, but they may rather be explained by the trapping of radiation between individual trees on a large scale and, on a much smaller scale, between the needles which take the place of leaves in most coniferous species. It is unlikely that the shape and disposition of these needles represent a special adaptation to the low light intensities which prevail for much of the year in regions where conifers thrive : but it is worth noting that the short-wave radiation absorbed annually by a pine forest in the north of Scotland may exceed that absorbed by pasture in the south of England because a smaller reflected component may more than compensate for a smaller incoming component.

4. Reflection as a function of wavelength

That plants with leaves of different shades of green should reflect almost the same fraction of incident radiation suggests that most of the reflection takes place outside the visible region of the spectrum and that the dependence of reflectivity $r(\lambda)$ on wavelength does not vary greatly among common species of similar type. This is confirmed by the work of Billings and Morris (1951) who measured the reflection coefficient of detached leaves with a spectrophotometer in the range 0·4-1·1 μ. For the majority of 20 species selected from widely different habitats reflectivity was relatively low (0·05-0·20) below 0·7 μ but increased abruptly to a constant value between 0·50 and 0·70 from 0·7 to 1·1 μ.

Billings and Morris do not give total reflection coefficients but these can be calculated from

$$\alpha = \int r(\lambda)\, I(\lambda)\, d\lambda \Big/ \int I(\lambda)\, d\lambda$$

where $I(\lambda)$ is the intensity of incident radiation. Integration can safely be restricted to between 0·4 and 1·1 μ because there is little solar radiation outside these wavelengths

and because for most vegetation $r(\lambda)$ drops below 0·1 beyond 2 μ (Gates and Tantraporn 1952). $I(\lambda)$ has been taken for a clear sky with zenith sun (Brunt 1939) and the integrals have been estimated by summing $r(\lambda)$. $I(\lambda)$ at 0·05 μ intervals. For four trees and shrubs growing in ' a thoroughly shaded and watered portion of the University of Nevada campus ' α ranges from 0·26 for *Euonymus europaea* (Spindle tree) to 0·21 for *Symorphicarpos albus* (Snowberry), both species having smooth, green leaves. Desert species have higher reflection coefficients achieved in two ways : (*i*) by a dense mat of fine hairs which increase reflectivity in the visible but not in the infra-red, e.g., *Eurotia lanata*, $\alpha = 0{\cdot}39$; (*ii*) by increased reflectivity in the infra-red but not in the visible, e.g., *Prunus andersonii*, $\alpha = 0{\cdot}35$. Fig. 3 (*i*) to (*iii*) gives $r(\lambda)$ and $r(\lambda)$. $I(\lambda)$ for three of these species.

For a white part of a *Pelargonium zonale* (geranium) leaf, $\alpha = 0{\cdot}44$ but for a green part $\alpha = 0{\cdot}28$. There is a marked difference in reflectivity in the visible region (Fig. 3, *iv*) but beyond 0·8 μ the two surfaces are almost identical. This implies that whereas reflection of solar radiation takes place at all wavelengths from cell walls and unpigmented tissue, chlorophyll and other pigments absorbing strongly in the visible and in the infra-red beyond 2 μ are transparent only between 0·8 and 1·1 μ. There may be a limit beyond which reflectivity is unchanged with increasing pigmentation but even before this is reached, perceptible changes in *visible* reflection have little effect on *total* reflection, dominated by the near infra-red component. Similarity of reflection coefficient for a wide range of agricultural crops suggests that differences in the cellular structure of their leaves do not produce significant differences in infra-red reflection below 1 μ.

Fig. 3 shows that the visual reflection coefficient of vegetation will normally be smaller than the total reflection coefficient and that the relation between the two depends on species. It is unfortunate that in the *Smithsonian Meteorological Tables* (1951) confusion is risked by the printing of both coefficients under the same head 'albedo', even though the type of instrument used is specified. Such confusion could be avoided by reserving the term 'albedo', literally 'whiteness', for the reflection coefficient in the visible range of the spectrum and by denoting the fraction of total short-wave radiation reflected by

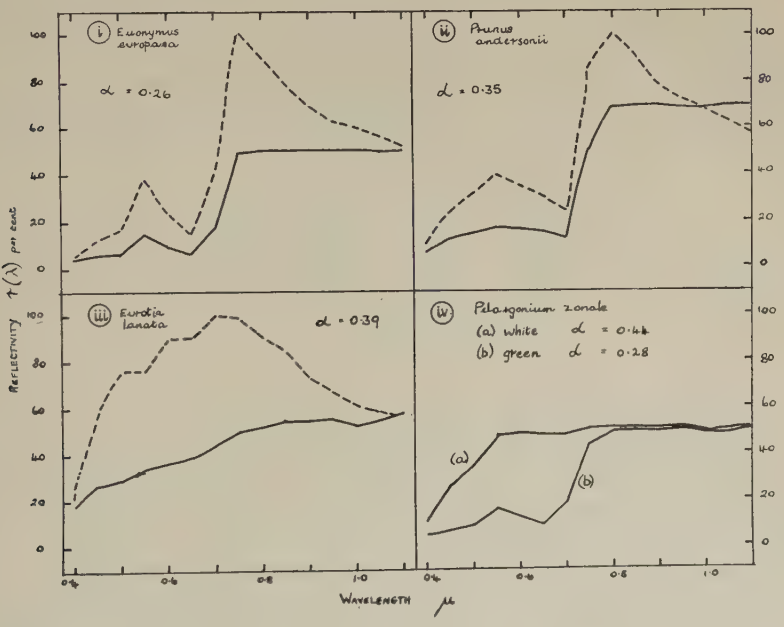

Figure 3. Reflectivity $r(\lambda)$ as a function of λ (full lines) and relative reflected energy $r(\lambda)\, I(\lambda)$ expressed as percentage of maximum value (dashed lines). $r(\lambda)$ redrawn from figures by Billings and Morris (1951). $I(\lambda)$ from Brunt (1939). (*i*) *Euonymus europaea*; (*ii*) *Prunus andersonii*; (*iii*) *Eurotia lanata*; (*iv*) *Pelargonium zonale* (a) white, (b) green.

' reflection coefficient '. By definition, instruments designed to measure ' visual ' albedo should have the same response as the human eye. Since the reflection of vegetation changes very rapidly with wavelength between 0·6 μ and 0·7 μ, comparatively small differences in instrumental response over this range can produce significant variations in observed ' visual ' albedo.

5. Conclusions

Many agricultural crops which completely shade the ground have reflection coefficients close to 0·26, and since large differences in surface temperature and hence long-wave emission are not to be expected for freely transpiring vegetation, differences in the net radiation received by different crops should be small. Other types of vegetation may absorb more radiation than crops, e.g., forests; or less radiation, e.g., desert species with white or very hairy leaves. Forests can therefore make more efficient use of available light, while desert plants can conserve water and reduce heating of the leaf above ambient air temperature.

Acknowledgments

Mr. G. Szeicz gave valuable help both with the observations and with Russian translations. Information on specific crops has been abstracted from the detailed records of colleagues at Rothamsted, and Dr. B. Orchard was responsible for the leaf-area determinations on sugar beet.

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A synoptic study of day-to-day changes of ozone over the British Isles

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(Manuscript received 20 November 1958; in revised form 28 April 1959)

Summary

Two months of European ozone data are analysed in relation to 24-hr trajectories determined from 100 mb contour charts. Significant correlations are obtained between vertical movements at 100 mb and non-advective and local changes of ozone, 100 mb temperature and 100 mb vorticity. These confirm the importance of horizontal convergence as a dynamic effect in the lower stratosphere.

Further correlations are presented between these changes and the accompanying N-S displacements in an attempt to relate short-period phenomena and the observed global distribution of ozone. The results do not conform to the simple models of the atmosphere considered here.

List of symbols

O = total (vertically integrated) ozone amount, measured in units 10^{-3} cm as determined by the absorption coefficients of Ny and Choong (1932).

T = 100 mb temperature, measured in °F.

ζ = 100 mb relative vorticity 10^{-5} sec^{-1}

l = Coriolis parameter ($\zeta + l$) in units 10^{-5} sec^{-1}

Thus ($\zeta + l$) is the 100 mb absolute vorticity.

ϕ = latitude.

Δ_t = local 24-hr change of a parameter at the end of a determined trajectory.

Δ_A = portion of the local change due to advection alone, and is equated to the initial difference between the values of the parameter at the beginning and end points of the trajectory.

Δ_N = non-advective change of the parameter along the trajectory.

Superscript 1 characterizes quantities measured at the end of a 24-hr 100 mb trajectory.

1. Introduction

This paper examines the dynamic causes of the day-to-day variations of total ozone which in middle latitudes are closely correlated with the variation of meteorological parameters in the lowest 20 km of the atmosphere. Normand (1953) showed that large ozone amounts usually accompany a low tropopause, a thin cold 500-300 mb layer, cyclonic curvature of the contour lines at 300 mb, and a warm lower stratosphere, while small ozone amounts accompany the reverse situation. These short-period ozone changes are most pronounced in spring when the average total ozone amount and its latitudinal gradient are largest, and are smallest in the autumn when a low roughly uniform total ozone amount is observed in the northern hemisphere.

The earlier estimates of the vertical distribution of the ozone placed most of the gas between the tropopause and 30 km; concentrations had a maximum in the region 20-25 km, decreasing sharply above the peak but gradually below it. The ozone below

25 km was thought not to be affected by photochemical activity, and day-to-day ozone changes were attributed mainly to dynamical effects. However, the finer details of the ozone distribution were not discernible from 'umkehr,' rocket and the usual balloon measurements, and examination of day-to-day ozone changes on the basis of simultaneous estimates of ozone concentration, wind and vertical motion was not possible.

In 1952 Kay (1953) obtained measurements from aircraft of the ozone concentration up to 50,000 ft (15 km) over southern England. These measurements did not substantiate the large amounts of ozone in the lower stratosphere given by other methods, and indicated no regular discontinuity of the ozone mixing ratio at the tropopause or at any level up to 50,000 ft. At the same time the network of ozone measuring instruments in north-west Europe and the Atlantic was providing sufficient data for a limited synoptic study, and daily soundings of wind and temperature extended into the stratosphere. Accordingly, we decided to examine day-to-day changes of total ozone measured in the autumn in conjunction with 100 mb (16 km) flow patterns on the assumption that most of the ozone occurred above this level, which was the highest for which reliable contour charts could be drawn on a 12-hourly basis. Subsequent ozone-sonde profiles show that such an assumption would not be valid for the spring, but that it is substantially correct in the autumn.

Our prime object was to relate measured ozone changes occurring along trajectories at 100 mb with vertical motion and other indices of dynamical activity which could be estimated from the contour charts. In particular, evidence was sought for the effect on the total ozone of horizontal and vertical divergence (which may be expressed alternatively as the effect of the vertical motions) occurring in the stratosphere, in terms of a relationship between the change of ozone along a trajectory and the movement of the lower boundary of the ozone column. Unfortunately, reliable trajectory data were not abundant, and correlations between parameters and their changes measured at fixed observing stations were also examined. We have to bear in mind, however, that advective effects also make contributions to these quantities, and may have obscured the particular effects we were seeking.

Our second objective was to discern some relationship between these same dynamical effects and the large-scale workings of the atmosphere which produce the observed mean distribution of ozone. Several earlier workers have constructed models of atmospheric behaviour to account for the available observations, but lack of detailed synoptic information on the upper atmosphere has denied these attempts an adequate basis. In the later sections of this paper the problem is approached by correlating the day-to-day advective and non-advective changes already mentioned with the local changes and latitude displacements which accompanied them, in the hope that definite trends would emerge, capable of straightforward interpretation.

2. Theoretical relationships examined in the study

As a basis to the analysis we require formal relationships between dynamical effects and changes of total ozone. To obtain these, some simplification of the complete picture of the processes involved is necessary, and we make the following assumptions :

(i) that solar radiation effects make no contribution to day-to-day changes of total ozone. This is consistent with the observed absence of any regular diurnal variation of total ozone.

(ii) that the effects of eddy and molecular diffusion are negligible compared with those of large-scale processes over a 24-hr period. This is consistent with the very stable temperature configuration of the lower stratosphere, and in any case no values of the eddy diffusion coefficient are available.

(iii) that the vertical shear of horizontal advection of ozone relative to the motion at 100 mb is negligible. We are obliged to make this assumption since we do not know the distribution of wind and ozone above 100 mb.

(iv) that the ozone is contained in a limited layer of unknown thickness, whose boundaries lie entirely between, but close to, the isobaric surfaces $p_1 = 100$ mb below and some higher level p_2. Thus the motions of the 100 mb surface are taken to be equal to those of the lower boundary of the ozone layer; but no contribution to the change of the total amount of ozone in the vertical column can come from vertical advection across the boundary since we prescribe that there shall be no ozone there.

On the basis of these assumptions and the usual dynamic properties of large-scale motions, we may make deductions as to the origin of the day-to-day changes of ozone as follows :

(*a*) *Local changes are due solely to advection by the 100 mb wind and to dynamic non-advective effects*

This proposition embodies assumptions (i) and (iii), and we may state it algebraically as

$$\Delta_t O = \Delta_A O + \Delta_N O$$

or, dividing by O^1, the total ozone at the end of a trajectory,

$$\frac{\Delta_t O}{O^1} = \frac{\Delta_A O}{O^1} + \frac{\Delta_N O}{O^1} \quad . \quad . \quad . \quad . \quad (1)$$

(*b*) *The dynamic non-advective changes are due to convergence or divergence*

This follows from assumptions (ii) and (iv) and is indeed obvious from physical considerations. Analysis shows that this non-advective change of total ozone is equal to the integral along the 100 mb trajectory of the vertically integrated isobaric convergence weighted with the ozone mixing ratio. We cannot calculate this weighted convergence since we do not know the vertical distribution of ozone, and we therefore make further restricting assumptions to derive expressions for subsequent trial.

(*c*) *Non-advective changes are related to the thickness of a layer of uniform mixing ratio*

Following assumption (iv), if the ozone layer has uniform mixing ratio μ then the total ozone O is given by $\mu(p_1 - p_2)$ and following a trajectory $\frac{\Delta_N O}{O^1} \simeq \frac{\Delta p_1 - \Delta p_2}{p_1 - p_2}$ if Δp_1 and Δp_2 are small.

If we neglect the effects of the upper boundary, i.e. assume that $p_2 \ll p_1$, $\Delta p_2 \ll \Delta p_1$, we obtain

$$\frac{\Delta_N O}{O^1} \simeq \frac{\Delta p_1}{p_1} \quad . \quad . \quad . \quad . \quad . \quad (2)$$

Thus for this special case of a uniform ozone layer, the proportional change of ozone is equal to the unweighted convergence above the lower boundary integrated along the trajectory.

Combining this with Eq. (1) we have

$$\frac{\Delta_t O}{O^1} \simeq \frac{\Delta p_1}{p_1} + \frac{\Delta_A O}{O^1} \quad . \quad . \quad . \quad . \quad (3)$$

and we assume that p_1 is near enough to the 100 mb level to be reasonably identified with it.

The movements of the lower boundary are thus related directly to proportional local changes of total ozone.

(*d*) *The vertical movements of the lower boundary are related to changes of its temperature*

The vertical movements of the lower boundary will be accompanied by adiabatic changes of its temperature, T, by means of which we can estimate the movements. Thus

$$\Delta p_1 \simeq \frac{1}{(\gamma_d - \gamma)_m} \Delta_N T \quad . \quad . \quad . \quad . \quad (4)$$

where $\gamma = \left[\frac{\partial T}{\partial p}\right]_{p=p_1}$ and $\gamma_d = \left[\left(\frac{\partial T}{\partial p}\right)_s\right]_{p=p_1}$ is the adiabatic lapse rate. $(\gamma_d - \gamma)_m$ represents a mean value along the trajectory.

By analogy with proposition (a) we have also

$$\Delta_t T = \Delta_N T + \Delta_A T \quad . \quad . \quad . \quad . \quad (5)$$

and so

$$\Delta_t T \simeq (\gamma_d - \gamma)_m \Delta p_1 + \Delta_A T. \quad . \quad . \quad . \quad (6)$$

(*e*) *The vertical movements of the lower boundary and changes of total ozone are related to changes of vorticity*

It is well known from the equation of continuity and the vorticity equation that the isobaric convergence at any level is approximately equal to the rate of change, following the motion, of the logarithm of the absolute isobaric vorticity. From this it may be shown that the movement of the lower ozone boundary, and the unweighted means of the convergence and of the logarithmic rate of change of vorticity above are all directly related to each other.

If we assume that the 100 mb level also represents the mean of higher levels, it follows that

$$\frac{\Delta p_1}{p_1} \simeq \frac{\Delta_N (\zeta + l)}{(\zeta + l)^1} \quad . \quad . \quad . \quad . \quad . \quad (7)$$

and from equation (2) that

$$\frac{\Delta_N O}{O^1} \simeq \frac{\Delta_N (\zeta + l)}{(\zeta + l)^1} \quad . \quad . \quad . \quad . \quad (8)$$

This is equivalent to the statement that

$$\frac{O}{(\zeta + l)} \text{ is constant along a 100 mb trajectory.} \quad . \quad . \quad (9)$$

Assuming (as for ozone and temperature) that local changes are due solely to large-scale advection and vertical motion we have

$$\Delta_t (\zeta + l) = \Delta_N (\zeta + l) + \Delta_A (\zeta + l).$$

On dividing by $(\zeta + l)^1$, the vorticity at the end of the trajectory, we have the relation among proportional vorticity changes

$$\frac{\Delta_t (\zeta + l)}{(\zeta + l)^1} = \frac{\Delta_N (\zeta + l)}{(\zeta + l)^1} + \frac{\Delta_A (\zeta + l)}{(\zeta + l)^1} \quad . \quad . \quad . \quad (10)$$

3. Data and techniques employed in numerical tests of the above relationships

To test the validity of these relationships we studied European and Atlantic data from 11 September to 10 October 1953. At this season the latitudinal gradient of monthly mean ozone amounts is small. Although the activity of the atmosphere is not usually pronounced in the autumn (in this instance the circulation pattern was broken up), the coverage of ozone observations is much better than in winter. This is due partly to there

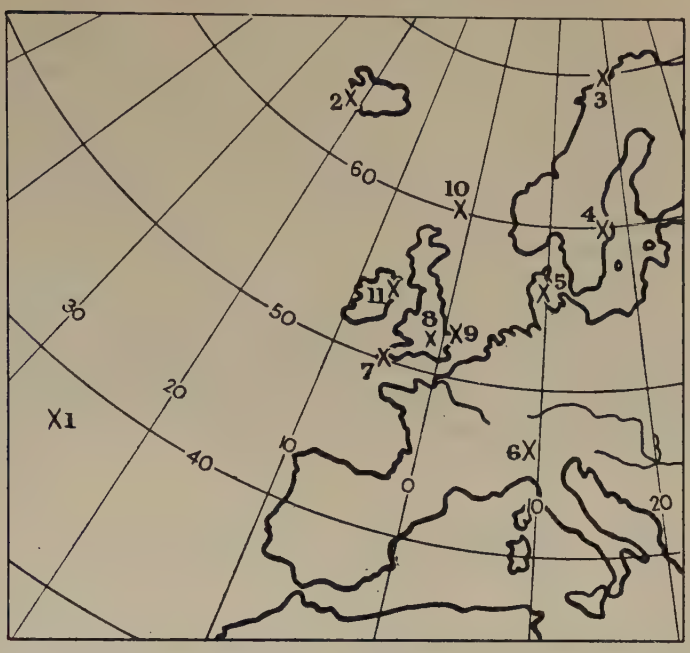

Figure 1. Ozone observations in Europe. 1. Santa Maria (Azores); 2. Reykjavik; 3. Tromsö; 4. Uppsala; 5. Aarhus; 6. Arosa; 7. Camborne; 8. Oxford; 9. Hemsby; 10. Lerwick; 11. Aldergrove.

being more sunshine for making observations at high latitudes, and partly to the reduced cloudiness at all latitudes. Total ozone amounts were available for most of the period at the stations shown in Fig. 1, and although the coverage leaves something to be desired it was possible to draw isopleths of total ozone on individual charts.

The daily isopleth patterns differed considerably from each other, and showed variations of total ozone with both latitude and longitude : even the mean distribution was not uniform. The individual charts showed large and small ozone values near troughs and ridges respectively, and closed isopleths occurred frequently, indicating non-advective agents of ozone change.

100 mb contour charts were drawn for 03 GMT and 15 GMT, allowance being made for the systematic difference between soundings at different times and by different instruments. The charts were constructed in accordance with the techniques of the Meteorological Office, Dunstable, and attention was given also to the daily German 96 mb charts.

We assumed that the vertical wind shear at 100 mb was negligible, and that the flow was in a steady state for the 12-hr periods centred on the time of each chart. Particle trajectories were identified accordingly with the contour lines, the large apparent ageostrophic component of 100 mb winds being ascribed to errors in the contour height observations (Johnson 1954). The vertical displacements were obtained by following, on a tephigram, the dry adiabat from the interpolated 100 mb temperature at the initial point of the trajectory to its intersection with the plotted radiosonde ascent at the terminal ozone station. Estimates of Δp_2 could not be made for want of data.

The vertical displacements amounted to 10 mb or more for only 7 of the 132 trajectories examined (1 mb per 24 hr is 0·05 cm sec^{-1} at 100 mb). The maximum 24-hr displacement was 16 mb (0·85 km) corresponding to a mean velocity of 0·8 cm sec^{-1}, and the r.m.s. displacement was 4 mb. These are magnitudes of the same order as those obtained by Miles (1952).

To determine the geostrophic relative vorticity at 100 mb, the usual approximation and finite difference techniques were used :

$$\zeta \simeq \frac{g}{l} \nabla^2 h = 2H \text{ in units } 10^{-5} \text{ sec}^{-1} \text{ at latitude } 50°$$

where $H = \sum_{i=1}^{4} h_i - 4h_0$, h being the contour height measured in hundreds of feet at the centre and extremities of a cross of semi-arm 200 naut. mi. The proportional vorticity changes exceeded 15 per cent for 41 of the 145 trajectories plotted and the r.m.s. change was about 13 per cent.

Substantiation of the theoretical relationships was sought by means of linear correlations between the quantities occurring in Eqs. (1) to (10). We attempted also to detect the downward eddy flux of ozone by correlating the vertical displacement Δp_1 with the mean ozone values along the trajectory $\bar{O}$, and with the terminal ozone value O^1.

In the later sections the importance of displacements towards or away from high latitudes, which for most of the year have more ozone than low latitudes, was examined by correlating the changes of ozone and temperature with the change of latitude, $\Delta\phi$, spanned by individual trajectories.

4. Errors of the observations

(*a*) *Total ozone*

The most reliable ozone measurements are usually those made on direct sunlight, which have a relative error of one per cent (Dobson and Meetham 1934). In the absence of direct sunlight, the ozone amount can be obtained by empirical relationships from measurements of the light scattered from the zenith blue sky or from a cloudy sky: cloud readings are thought to be the least reliable. Errors in the zenith blue and cloud measurements have been discussed by Tønsberg and Olsen (1944) and by Langlo (1952). The estimated maximum error in a single zenith blue observation made at noon is 0·01 cm in October, and an additional error of 0·01 cm may arise in cloud measurements. Clearly, such errors could yield spurious changes in the total ozone amounts, which averaged 0·210 cm in this study. However, no consistent relationship was discerned between the different kinds of measurement at different stations, and equal weight was given to all the observations at each station in assessing the daily ozone field. Thus errors in individual observations should have been smoothed to some extent.

(*b*) *Temperature and vertical displacement*

The random error of temperature for individual British soundings was thought not to exceed 2 to 3°F at 100 mb (Air Ministry 1954), and Bannon and Gilchrist (1953) quote a standard deviation of 2°F for casual errors of an older British sonde in the stratosphere. These errors were probably offset to some extent by smoothing the isotherms from which 100 mb temperatures were interpolated.

The probable error in Δp_1 as shown by temperature changes, due to a probable error by the sounding instrument of 1°F in observed temperatures in the lower stratosphere, is about 1·5 mb (Sawyer 1954). The estimated r.m.s. error in Δp_1 is about 2 mb, one half the r.m.s. of the displacements calculated for this study. Thus the r.m.s. of the displacements after allowance has been made for such error is 3·5 mb.

Persistent error in the estimation of wind speeds (such as arises from neglecting the curvature of the earth), and hence of $\Delta_A T$, can give false estimates of vertical motion when particle trajectories traverse the isotherms. According to Gustafson's data (1954), a 24-hr 100 mb trajectory could be overdrawn by 25 naut. mi. if the mean wind up to 100 mb were 30 kt and 100 naut. mi. if the mean wind were 60 kt. The errors from this source in the temperatures interpolated at the beginning of the trajectories were probably 2°F or less; the corresponding errors in Δp_1 are about 1 mb per 1°F.

(c) *Vorticity*

Johnson (1953, 1954) has shown that although 100 mb contour height observations contain serious errors, the 100 mb wind can be measured with reasonable precision. By expressing the finite difference formula in terms of wind gradients, the standard error in the geostrophic vorticity at 100 mb in the vicinity of the British Isles, due to errors in the winds used in drawing the contour charts, was estimated to be about ten per cent of the Coriolis parameter (Martin 1958). While this is too small to obscure the vorticity pattern at any point, the standard error in vorticity changes is as large as the r.m.s. of the changes actually measured.

5. Results concerning vertical motion in the stratosphere

The results obtained are given in statistical form in Table 1 : all quantities relate to 15 GMT. One set was obtained from only the most reliable data; the other set from all the data available.

TABLE 1. Correlations concerning vertical motion in the stratosphere

x and y denote the variables correlated. N is the number of pairs of variables used to calculate the correlation coefficient, r. The significance, S, is given as the odds against the value of r occurring by random sampling from an uncorrelated population : — denotes odds of less than ten to one. The equations between y and x are the mean regressions; these have been made non-dimensional by introducing $|O|$ to denote the average of O for the period studied, and L to denote l at latitude 50°.

			Most reliable				Total			
	x	y	N	r	S	Linear regressions	N	r	S	Linear regressions
(1)	$\frac{\Delta_N O}{O^1}$	Δp_1	62	+ 0·58	$10^3 : 1$	$y = 0{\cdot}41x + 1{\cdot}41$	100	+ 0·36	$10^3 : 1$	$y = 0{\cdot}24x + 0{\cdot}09$
(2)	$\frac{\Delta_t O}{O^1}$	Δp_1	81	+ 0·35	$10^3 : 1$	$y = 0{\cdot}21x + 0{\cdot}15$	109	+ 0·25	$10^2 : 1$	$y = 0{\cdot}16x - 0{\cdot}07$
(3)	$\Delta_t T$	Δp_1	72	+ 0·66	$10^3 : 1$	$y = 0{\cdot}45x - 0{\cdot}14$	109	+ 0·65	$10^3 : 1$	$y = 0{\cdot}41x - 0{\cdot}04$
(4)	$\frac{\Delta_N(\zeta+l)}{(\zeta+l)^1}$	Δp_1	62	+ 0·02	—		100	+ 0·06	—	
(5)	$\frac{\Delta_N(\zeta+l)}{(\zeta+1)^1}$	$\frac{\Delta_N O}{O^1}$	62	− 0·20	—		100	− 0·16	—	
(6)	$\overline{O}$	Δp_1	62	0	—		100	0	—	
(7)	$\frac{O}{\|O\|}$	$\frac{\zeta+l}{L}$	85	+ 0·46	$10^2 : 1$	$y = 1{\cdot}1x$	126	+ 0·48	$10^3 : 1$	$y = 1{\cdot}1x$
(8)	$\frac{O}{\|O\|}$	T	85	+ 0·55	$10^3 : 1$	$y = 67{\cdot}2x - 130{\cdot}0$	126	+ 0·57	$10^3 : 1$	$y = 71{\cdot}4x - 135{\cdot}0$
(9)	$\frac{\zeta+l}{L}$	T	85	+ 0·74	$10^3 : 1$	$y = 38{\cdot}9x - 104{\cdot}8$	126	+ 0·71	$10^3 : 1$	$y = 37{\cdot}7 - 103{\cdot}7$
(10)	$\frac{\Delta_t O}{O^1}$	$\frac{\Delta_t(\zeta+l)}{(\zeta+l)^1}$	104	+ 0·16	10 : 1	$y = 0{\cdot}38x - 0{\cdot}86$	110	+ 0·16	10 : 1	$y = 0{\cdot}42x - 1{\cdot}03$
(11)	$\frac{\Delta_t O}{O^1}$	$\Delta_t T$	58	+ 0·25	20 : 1	$y = 0{\cdot}20x + 0{\cdot}79$	110	+ 0·35	$10^3 : 1$	$y = 0{\cdot}34x - 0{\cdot}13$
(12)	$\frac{\Delta_t(\zeta+l)}{(\zeta+l)^1}$	$\Delta_t T$	75	+ 0·52	$10^3 : 1$	$y = 0{\cdot}18x + 0{\cdot}79$	110	+ 0·34	$10^3 : 1$	$y = 0{\cdot}13x - 0{\cdot}11$

Bannon and Gilchrist's discussion (1953) of persistence in upper-air variables measured at a fixed station suggests that the number of independent observations is N for the differenced quantities occurring, and $\frac{1}{2}N$ for the un-differenced quantities.

The correlations obtained will now be discussed with reference to the derivations of § 2.

(a) Relations between convergence in the ozone layer and changes of total ozone

Results (1) and (2) of Table 1 relate to Eqs. (2) and (3) derived in § 2 (c). The high significance of the correlations is gratifying since the ozone changes depend on the whole field of vertical motions above 100 mb and, in the case of the local changes, the contribution of advection. The departure of the correlation coefficients from unity is presumed to reflect the simplifications of the present discussion, for the proportion of variance ascribable to the linear relationships is small. An estimate of the 100 mb vertical displacement which contributes to a particular ozone change is given by the regression equations, but the size of the correlation coefficients indicates that considerable departures from these linear relationships occurred.

(b) Vertical motion and local temperature changes at 100 mb

Result (3) shows a significant correlation between two of the quantities occurring in Eq. (6) of § 2 (d), and we conclude that the effects of advection did not obliterate the effects of non-advective processes on local changes of temperature. The correlation coefficient is larger than that for ozone, possibly because the non-advective changes of temperature depend directly on the vertical motion at 100 mb, but possibly also because advection and non-advective processes make unequal contributions to the changes of the different parameters. (The possibility of this result arising from the systematic underestimation of $\Delta_A T$ has been discussed in § 4 (b)).

(c) Relations of movements of the lower boundary of the ozone layer and changes of total ozone with changes of vorticity

Relations (7) and (8) of § 2 (e) are not supported by correlations (4) and (5) of Table 1, and it seems that the change of vorticity at 100 mb as derived from the charts is a poor index of the vertical motions occurring at that level and above. These results are surprising in view of correlation (7) between ozone and vorticity and the similar results of earlier workers, but it may be that non-advective changes of vorticity at any level depend largely on the vertical derivative of the vertical motions at the level, and not directly on the motions themselves. Alternatively, the low correlations (4) and (5) could easily be due to inaccuracy in calculating vorticity changes from contour charts, as discussed in § 4 (c).

(d) Vertical eddy flux of ozone

The negligible correlation (6) between the vertical displacement from 100 mb along a trajectory and the mean of the ozone amounts at the beginning and the end of the trajectory offers no immediate evidence for a vertical eddy flux of ozone by the motions of synoptic scale. However, reliable conclusions on the vertical eddy flux require correlations between the vertical velocity and the ozone mixing ratio at individual levels (cf. Reed 1953).

(e) Relations between total ozone, and temperature and vorticity in the lower stratosphere

The highly significant correlations (7), (8) and (9) provide further confirmation of the observations of Dobson and Normand that near the British Isles troughs in the lower stratosphere tend to be regions of high temperature and large ozone amounts. Conversely, ridges are cold and are regions of small ozone amount.

The correlations (10), (11) and (12) between local changes of ozone, vorticity and temperature show the preceding relationships with reduced coefficients, possibly due to the difficulty of detecting small differences between quantities which contain errors.

(*f*) *Summary of these results*

The most impressive correlations are those between temperature and vorticity, between their local changes, and between local temperature changes and vertical motion. We conclude that the dynamical processes which affect temperature are more susceptible of quantitative assessment than the processes affecting ozone.

In other respects the results in Table 1 are more or less what was to be expected on the basis of the hypotheses suggested in § 2. The vertical displacements measured in the lower stratosphere are a significant index of the non-advective processes which cause changes of ozone, while correlations with vorticity changes were disappointing. However, even for the best correlations involving ozone the proportion of variance accounted for is small. There was no indication of a downward eddy flux of ozone by the motions of synoptic scale.

6. The relation between day-to-day advective and non-advective changes and the mean distribution of ozone and temperature

Our discussion so far has concerned merely the detection of non-advective processes occurring along restricted trajectories in the stratosphere. A more complicated problem is to understand how such short-period changes are related to the large-scale workings of the atmosphere.

The stratosphere appears to have operating within it two distinct mechanisms which have been variously invoked by earlier workers to explain some feature of the day-to-day fluctuations of total ozone. Firstly we have a very clear and well-substantiated average distribution of temperature and ozone, in which the total ozone amounts and stratospheric temperatures are greater in high latitudes than in low. Disturbances of this average regime would obviously produce local day-to-day variations of these parameters.

Secondly we have superimposed upon this average distribution (and perhaps indeed helping to maintain it) a stratospheric wave-system associated with the well-known long waves in the westerlies, in which the temperature, vorticity and ozone at fixed observatories are all strongly correlated. Examination of the temperature changes occurring along streamlines in the lower stratosphere indicates that subsidence occurs along southward trajectories from cold ridges to warm troughs, and vice versa (Air Weather Service 1954). These vertical motions are presumably accompanied by non-advective changes of ozone like those discussed above. Advective effects must also occur in association with these displacements, and it is reasonable to assume that advection operates on a distribution of high and low ozone amounts in the troughs and ridges respectively. Although adequate observations of ozone distribution along streamlines are not yet available to test this assumption, it is consistent with the high local correlations so far obtained between ozone, temperature and vorticity. The local changes which result from this advection pattern are obviously opposite in sign to those produced by advection through the mean pattern.

A clear understanding of the roles of these different processes can be achieved only by examining actual advective and non-advective components of local changes, together with the extent of the associated north-south air movements. Accordingly we now return to our data and examine it with reference to two models which incorporate the possible alternative effects just described. In Model A northward streamlines advect from warm ozone-rich troughs to cold ozone-poor ridges and vice versa. Model B, on the other hand, supposes that advection operates day by day on the poleward increases of ozone amount which is usually clearly shown in monthly means; and in fact this was the earliest model used to explain local ozone changes.

We associate with both models the same pattern of vertical movements, namely air moving south subsides (as from ridge to trough), while air moving north ascends. The associated non-advective changes of temperature and ozone are deduced from Table 1. The signs of the correlations which would result from these hypothetical changes are presented in Table 2 together with those actually obtained from our data.

TABLE 2. COMPARISON OF HYPOTHETICAL CORRELATIONS WITH THOSE ACTUALLY OBTAINED

			Model A	Model B	Actual
(13)	Δp	$\Delta\phi$	—	—	+
(14)	$\Delta_N O$	$\Delta\phi$	—	—	(+)
(15)	$\Delta_A O$	$\Delta\phi$	+	—	+
(16)	$\Delta_t O$	$\Delta\phi$	?	—	+
(17)	$\Delta_N T$	$\Delta\phi$	—	—	+
(18)	$\Delta_A T$	$\Delta\phi$	+	—	(+)
(19)	$\Delta_t T$	$\Delta\phi$	?	—	+
(20)	$\Delta_N O$	$\Delta_A O$	—	+	—
(21)	$\Delta_N O$	$\Delta_t O$	?	+	+
(22)	$\Delta_A O$	$\Delta_t O$	?	+	+
(23)	$\Delta_N T$	$\Delta_A T$	—	+	(—)
(24)	$\Delta_N T$	$\Delta_t T$	?	+	+
(25)	$\Delta_A T$	$\Delta_t T$	?	+	+

? indicates that the correlation is not predictable from the model. Results in brackets are not significant at the 10 per cent level.

7. RESULTS RELATING TO THE LARGE-SCALE BEHAVIOUR OF THE ATMOSPHERE

Our observations indicate that temperature, ozone and vorticity all tended to increase with northward and to decrease with southward trajectories. The signs of the local changes were determined sometimes by advection and sometimes by non-advective processes, and these effects were usually opposed. The actual magnitudes involved revealed no clear inter-relationship.

It is apparent that our results taken together are not compatible with either of the two most obvious models available. The most striking discrepancy concerns the correlation predicted between vertical motion and latitude displacement (Table 2 (13)) although this is usually regarded as a well established result. However, even if we declare the period studied to be abnormal in comparison with the usual pattern of vertical motion through the trough-ridge systems of the upper westerlies, it is still not possible to synthesize all our results. For example, Model A with reversed vertical motion (descent from trough to ridge) would not account for the negative correlations (20) and (23) between advective and non-advective changes, although advection and the effects of vertical motion tend to cancel each other in numerous synoptic situations.

No simple resolution of these inconsistencies is apparent. A possible conclusion is that our data relate to trajectories from more than one homogeneous and self-consistent group, each of which would reflect some particular feature of the aggregate results. In a more extensive study some preliminary classification of the trajectories (e.g., with regard to circulation patterns) might be attempted in the hope that unambiguous conclusions would be obtained about each group of trajectories. This approach was considered here, but was rejected on account of the difficulty of assigning a sufficient number of the trajectories available to well-defined classes.

8. FINAL CONCLUSIONS

The first part of this paper shows, as would be expected on physical grounds, that vertical displacements measured along 24-hr trajectories in the lower stratosphere are very significantly correlated with the proportional changes of total ozone amount which occur along the trajectories. Alternatively, the unweighted isobaric convergence above 100 mb is a significant index of the vertical field of convergence weighted at each level with the

ozone mixing ratio. However, despite their statistical significance, the magnitude of the correlation coefficients leaves much of the variance unexplained. Detailed information on the vertical distribution of both convergence and ozone is necessary to improve our understanding of the processes involved, and such information is not yet obtainable.

Nonetheless, tentative conclusions on the distribution of vertical velocity in the upper stratosphere are available (Air Weather Service 1953). The regime of vertical motion in the lower stratosphere appears to weaken from 200 mb up to 80 mb or 60 mb and occurs with reversed sign at greater heights – possibly at 50 mb, and more clearly at 25 mb. Such a difference of sign of the vertical displacements at the upper and lower boundaries of the ozone layer would increase the unweighted mean convergence in the ozone region from the value indicated by the displacement of the lower boundary alone, and more satisfactory correlations should arise from even this much extra data.

Relations between non-advective changes of ozone and vorticity were rather poor and a disappointing return for the labour involved, probably owing to the difficulty of measuring vorticity to a sufficient accuracy. As observed by previous workers, regions of high absolute vorticity were characterised by high 100 mb temperatures and large ozone amounts; opposite effects characterised regions of low vorticity.

The results of the later sections concern the global distribution of ozone. In the period studied, subsidence tended to occur along northward trajectories, contrary to our expectations, and at the end points of such trajectories local changes of total ozone, 100 mb vorticity and 100 mb temperature tended to be positive. Although the local changes were positively correlated with both the advective and non-advective components, these components were negatively correlated with each other. The results show a high degree of internal consistency, but taken together they are not explicable in terms of the simple models invoked by previous workers, and the need for further information on the interrelation of advection and other dynamical processes is obvious.

Acknowledgments

We thank the Director-General of the Meteorological Office for his encouragement of this work, and Mr J. S. Sawyer for providing facilities at the Meteorological Office, Dunstable. The interest of Professor G. M. B. Dobson, F.R.S., and Sir Charles Normand is much appreciated; we are indebted to them and to others responsible for maintaining ozone stations for permission to use their unpublished observations.

The paper is based on work carried out on the General Circulation Project of the Department of Meteorology of the Massachusetts Institute of Technology. We thank Professor V. P. Starr for his hospitality and for the provision of facilities under the sponsorship of the Geophysics Research Directorate of the Air Force Cambridge Research Center.

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551.521.3

The role of water particles in the atmospheric transmission of infra-red radiation

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(Manuscript received 16 January 1959; in revised form 25 June 1959)

SUMMARY

The attenuation of infra-red radiation by scattering on atmospheric haze particles, in the form of water droplets, is discussed in some detail. The spectrum is divided into two regions, the near infra-red to λ 2·25 μ, where the refractive index of water may be assumed constant and real, and the region 2·25 to 12 μ, where the index is complex and varies with the wavelength. This paper is concerned with the first region. It is shown that certain haze models, involving a continuous drop-size distribution and previously used in explaining the attenuation of visible wavelengths and the aureole around the sun, adequately explain the window transmission measured by Gebbie and co-authors (1951) along a horizontal path. The observed optical thickness of the whole atmosphere with respect to near infra-red sunlight is also explained by means of these models.

The infra-red attenuation spectrum appears as a logical extension of the visible one and represents the continuous background upon which gas line and wing absorptions must be superposed. The magnitude of the latter can thus be assessed with more confidence.

1. INTRODUCTION

The recent development of refined techniques in the detection and measurement of infra-red radiation (Smith, Jones and Chasmar 1957) makes it possible to get a better picture of the atmospheric transmission, absorption, scattering, and emission of this radiation. Past approaches to the study of infra-red transmission and absorption in the atmosphere generally were concerned with the effect of various molecular lines and bands of water vapour, carbon dioxide, ozone and other gaseous constituents. The effect of the continuous background arising from the scattering of infra-red by haze particles and cloud droplets, has not received sufficient attention.

The purpose of the present discussion is to examine the effect of such scattering on the atmospheric extinction of infra-red. We base our discussion on the assumption that haze particles are mainly spherical water droplets, a hypothesis which has proved quite adequate in explaining the extinction and scattering of visible radiations (Deirmendjian 1957, 1959). In this paper we limit ourselves to that part of the infra-red spectrum (λ 0·8 to 2·25 μ) where liquid water has very weak or no absorption bands, and the index of refraction may be represented by a constant real number. In a further paper we shall extend the discussion to a wavelength of 14 μ, in which case the index of refraction is a complex number, both the real and imaginary parts of which vary considerably with the wavelength.

2. THE HORIZONTAL ATTENUATION BY SCATTERING ON HAZE PARTICLES

The imaginary part of the complex index of refraction for water is shown in Fig. 1 (top curve marked n_w'), adapted from the best available measurements (Centeno 1941; Curcio and Petty 1951), and plotted logarithmically against the wavelength. It is seen that, between λ 1 μ and 20 μ, this quantity varies by about six orders of magnitude, with several regions of relative maxima and minima. An equivalent quantity for water vapour is indicated by the lower curve marked n_v', such that $4\pi n_v'/\lambda$ gives the absorption coefficient

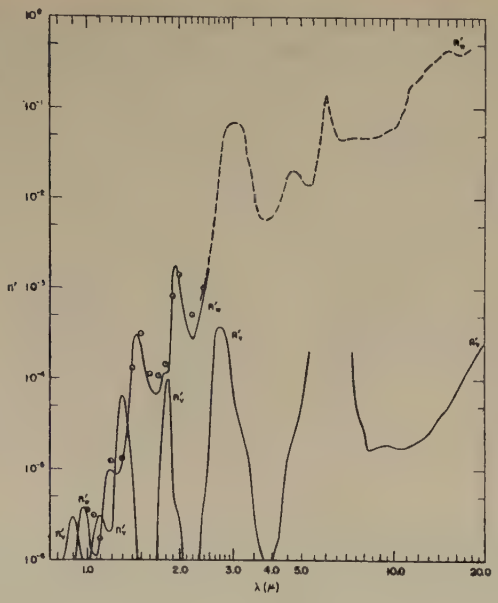

Figure 1. Spectrum of the imaginary part of the refractive index of liquid water (n_w') and of water vapour (n_v') at the same density.

corresponding to unit depth of precipitable water vapour. A comparison of the two curves in Fig. 1 shows at a glance the differences in the absorption spectrum of the two phases of water substance. The 'window' regions do not always coincide exactly. It should be noted that there is no general agreement about the effect of the various water-vapour bands, lines and wings on the monochromatic absorption coefficient in the infra-red spectrum.

A comparison of the scattering cross-section for dielectric spheres, computed according to the Mie theory (Penndorf 1956), with the extinction cross-section for slightly absorbing spheres with the same index (Johnson, Eldridge and Terrel 1954; Johnson and Terrel 1955), shows that the absorption may be neglected provided $n_w' < 0{\cdot}002$. The error in the extinction cross-section is less than $\pm$ 6 per cent for a given size of particle, and even smaller when several sizes are considered. Therefore, for infra-red radiation with $\lambda < 2{\cdot}25\ \mu$, it is sufficient to consider dielectric spheres with a refractive index $m = 1{\cdot}33$.

The important consideration in estimating the haze attenuation is the size-distribution spectrum of the aerosol particles. Recalling the variation of the scattering cross-section as a function of the relative size parameter $\alpha = \dfrac{2\pi a}{\lambda}$, where a is the actual radius of the particle, it is easy to show that the contribution of different sizes and concentrations depends very much on the portion of the curve which must be used. Assuming a *continuous* size distribution spectrum, as proposed by Sekera (1957), and spherical water particles only, the haze or Mie attenuation coefficient $\beta_M(0)$, at sea level, is given by the integral

$$\beta_M(0) = \text{Const} \times \int_{a_1}^{a_2} n(a)\, \pi a^2 K(\alpha, m)\, da \quad . \quad . \quad (1)$$

where $n(a)$ is the size-distribution function of the particles in numbers per cm^3 and per unit increment in the radius, a_1 and a_2 represent the smallest and largest radii encountered, and the constant factor is determined by the total number of particles. The function $K(\alpha, m)$ is the ratio of the scattering cross-section to the geometrical cross-section.

In the model size-distributions already used for the visible spectrum, the size range was divided into two or three parts according to the appropriate power law in each case. Thus, following the same designation as in a previous paper (Deirmendjian 1959) we have

$$\text{Model A} \quad \left\{ \begin{array}{lll} n = 1352{\cdot}8 & \text{cm}^{-3}\,\mu^{-1}; & a_1 \leqq a \leqq a_0 \\ n(a) = 42{\cdot}78\, a^{-1{\cdot}5} & \text{cm}^{-3}\,\mu^{-1}; & a_0 \leqq a \leqq a_c \\ n(a) = 15{\cdot}125\, a^{-3} & \text{cm}^{-3}\,\mu^{-1}; & a_c \leqq a \leqq a_2 \end{array} \right\} \quad . \quad . \quad (2)$$

$$\text{Model B}' \quad \left\{ \begin{array}{lll} n = 1964{\cdot}4 & \text{cm}^{-3}\,\mu^{-1}; & a_1 \leqq a \leqq a_0 \\ n(a) = 62{\cdot}12\, a^{-1{\cdot}5} & \text{cm}^{-3}\,\mu^{-1}; & a_0 \leqq a \leqq a_c \\ n(a) = 4{\cdot}502\, a^{-4} & \text{cm}^{-3}\,\mu^{-1}; & a_c \leqq a \leqq a_2 \end{array} \right\} \quad . \quad . \quad (3)$$

$$\text{Model C} \quad \left\{ \begin{array}{lll} n = 2{\cdot}251 \times 10^4 & \text{cm}^{-3}\,\mu^{-1}; & a_1 \leqq a \leqq a_0 \\ n(a) = 2{\cdot}251\, a^{-4} & \text{cm}^{-3}\,\mu^{-1}; & a_0 \leqq a \leqq a_2 \end{array} \right\} \quad . \quad . \quad (4)$$

The values of a_1 and a_0 were fixed at 0·03 μ and 0·10 μ respectively, for all three models, and a_c was chosen 0·50 μ for Model A and 0·35 μ for Model B′. These particular choices are not fortuitous but are based on actual counts and they are designed to represent various conditions such as turbulent marine air (Model A), coastal air (Model B′) and continental air (Model C). They involve no dominant or characteristic single sizes.

In order to evaluate the integral in Eq. (1), a change of variable is introduced by using $a = \frac{\alpha}{k}$, $da = \frac{d\alpha}{k}$, and this integral in general may be written as the sum of three terms,

$$\beta_M(0) = C_1 K_1 + C_2 K_2 + C_3 K_3 \quad . \quad . \quad . \quad (5)$$

where, for example, in the case of Model A, we have

$$\left. \begin{array}{lll} C_1 = 1352{\cdot}8\, \pi k^{-3}, & K_1 = \displaystyle\int_{\alpha_2}^{\alpha_0} \alpha^2 K(\alpha, m)\, d\alpha \\ C_2 = 42{\cdot}78\, \pi k^{-1{\cdot}5}, & K_2 = \displaystyle\int_{\alpha_0}^{\alpha_c} \sqrt{\alpha}\, K(\alpha, m)\, d\alpha \\ C_3 = 15{\cdot}125\, \pi, & K_3 = \displaystyle\int_{\alpha_c}^{\alpha_2} \alpha^{-1} K(\alpha, m)\, d\alpha \end{array} \right\} \quad . \quad . \quad (6)$$

We note that the value of $\beta_M(0)$ will depend on the wavelength since both the constants C_1, C_2, and the limits of integration in K_1, K_2 and K_3 depend on $k = 2\pi/\lambda$. Otherwise these integrals are functions of the single variable α, since m is assumed constant.

The function $K(\alpha, m)$ is given in the exact theory by a series which is laborious to compute, especially for large values of α, and cannot readily be integrated. The analytical approximation for the case $m \to 1$, first derived by van de Hulst in 1946 (see van de Hulst 1957), can be integrated in closed form *between finite limits of integration* with fairly good results (see Appendix). However, in the present instance we have resorted to numerical integration of all three integrals in Eq. (6) for the sake of uniformity and accuracy, taking advantage of the excellent computations of Penndorf (1956). The upper limit a_2 of the particle radius was initially fixed at 2 μ and the computations were carried out for the wavelengths $\lambda = 0{\cdot}625$, 0·800, 1·00, 1·65, 2·25 and 3·50 μ. These were repeated by increasing the value of a_2 to 4·48 μ, in order to find the effect of the larger particles on the attenuation coefficient. The results are summarized in Table 1 below, which also includes the corresponding molecular or pure-air attenuation coefficient (Deirmendjian 1955) for comparison.

TABLE 1. NEAR INFRA-RED ATTENUATION COEFFICIENTS $\beta\,(KM^{-1})$ AS AT SEA LEVEL

Model	$\lambda\,(\mu)$ 0·625	0·800	1·00	1·65	2·25	3·50
	$\beta_R\,(0)$ for molecular air, $2{\cdot}57 \times 10^{19}$ molecules per cm^3					
Rayleigh	$6{\cdot}62 \times 10^{-4}$	$2{\cdot}42 \times 10^{-3}$	$0{\cdot}99 \times 10^{-3}$	$0{\cdot}13 \times 10^{-3}$	$0{\cdot}037 \times 10^{-3}$	$0{\cdot}007 \times 10^{-3}\ km^{-1}$
	$\beta_M\,(0)$ for haze particles only, max. radius $2{\cdot}0\ \mu$					
Model A	0·245	0·233	0·219	0·174	0·119	$0{\cdot}044\ km^{-1}$
Model B′	0·127	0·107	0·0865	0·0458	0·0307	$0{\cdot}0109\ km^{-1}$
Model C	0·0822	0·0628	0·0488	0·0262	0·0156	$0{\cdot}0054\ km^{-1}$
	$\beta_M\,(0)$ for haze particles only, max. radius $4{\cdot}48\ \mu$					
Model A	0·328	0·321	0·308	0·268	0·231	$0{\cdot}184\ km^{-1}$
Model B′	0·136	0·116	0·0957	0·0555	0·0429	$0{\cdot}0252\ km^{-1}$
Model C	0·0864	0·0674	0·0534	0·0311	0·0217	$0{\cdot}0126\ km^{-1}$

3. COMPARISON WITH THE OPTICAL THICKNESS OF THE ATMOSPHERE

In order to find the monochromatic normal optical thickness for the whole atmosphere, the attenuation coefficients $\beta_R\,(0)$ of the molecular atmosphere, and $\beta_M\,(0)$ must be integrated with respect to the height h above the ground. If an exponential stratification of haze particles is assumed, i.e., if $\beta_M\,(h) = e^{-h/H_M}\,\beta_M\,(0)$, then this integration reduces to

$$\tau_M\,(0) = H_M\,\beta_M\,(0) \quad . \quad . \quad . \quad . \quad (7)$$

where $\tau_M\,(0)$ is the haze contribution, and H_M is the scale height of the haze component (taken as 0·98 km according to Penndorf 1954). The molecular contribution $\tau_R\,(0)$ has been computed by various authors (Deirmendjian 1955; Penndorf 1957) and can also be estimated without much error from $\beta_R\,(0)$ by using the scale height H_R of a standard atmosphere, say 8·4344 km, according to the ARDC model (Minzner and Ripley 1956).

The near and intermediate infra-red optical thickness of the turbid atmosphere in the range $0{\cdot}6 \leqq \lambda \leqq 3{\cdot}5\ \mu$, (excluding the effects of water vapour) computed according to some of the models, is shown in Fig. 2. The lowest straight line represents the molecular case (Rayleigh atmosphere) and the remaining full curves correspond to the various models (including the molecular effect) labelled according to Table 1. The open dots connected by broken lines represent the actual optical thickness at Miami, Florida, as deduced from observations of the Smithsonian Astrophysical Observatory (Aldrich and Hoover 1954). Observation conditions on 6 Feb. and 17 April 1948, were reported ' fair ' with 32·1 and 35·5 mm of precipitable water vapour respectively and ' poor ' on 11 June 1948, with 41·4 mm of water vapour. (The method of reduction used by the Smithsonian involves a ' smoothing over ' of the bolograph traces in the regions of H_2O and CO_2 band absorptions, so that only the haze effect is reflected in the reported transmissions). Although the Smithsonian data do not go beyond $\lambda\,1{\cdot}24\ \mu$, there is general agreement in magnitude and in slope between these and the theoretical curves for Models B′ and C. The increasing deviations from the theoretical values in the vicinity of $\lambda\,0{\cdot}6\ \mu$ can be attributed to the ozone absorption in the Chappuis bands. It is clear from Fig. 2 that neither Rayleigh scattering nor water-vapour absorption or a combination of both can explain the observed attenuation in the near infra-red.

According to the exponential stratification, the haze component $\tau_M\,(h)$ of the optical thickness drops to $\frac{1}{10}$ of its sea-level value at an altitude of 2·256 km. The corresponding curves (full lines) are plotted in Fig. 3 for each model, showing both types of size ranges, with maximum radii of $2\ \mu$ and $4{\cdot}48\ \mu$ respectively. These indicate the importance of a mere increase in the upper limit of the particle sizes, *without modifying the size concentration or the distribution law*, in both the magnitude of the haze attenuation and its spectrum

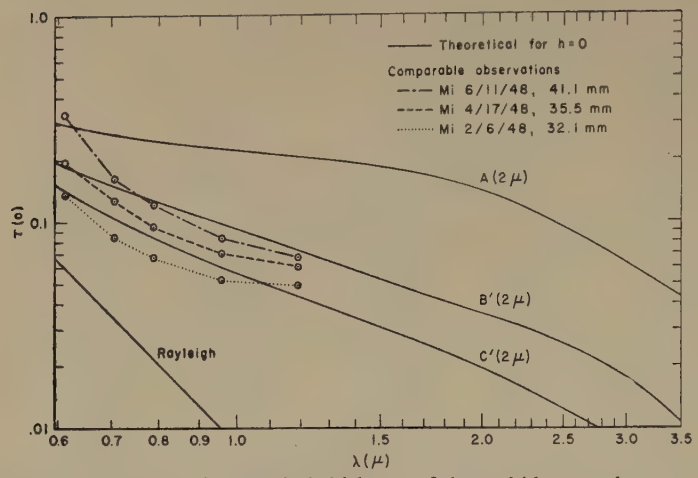

Figure 2. Near infra-red spectrum of the optical thickness of the turbid atmosphere at sea level, assuming water-coated aerosol particles.

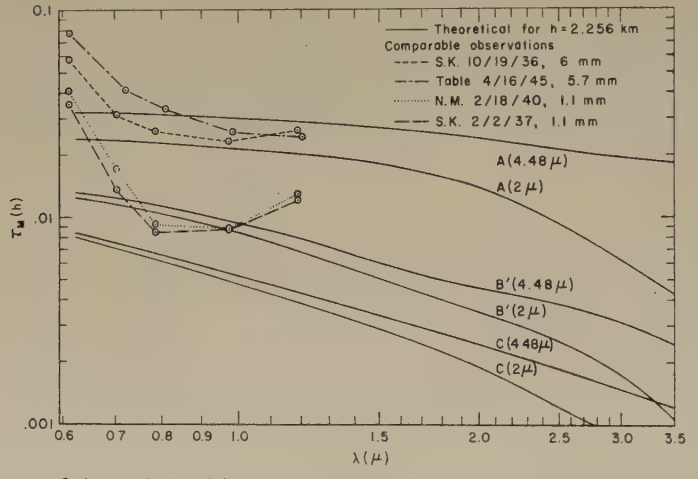

Figure 3. Spectrum of the optical thickness of the haze atmosphere only above a level of 2·256 km, and comparable measurements.

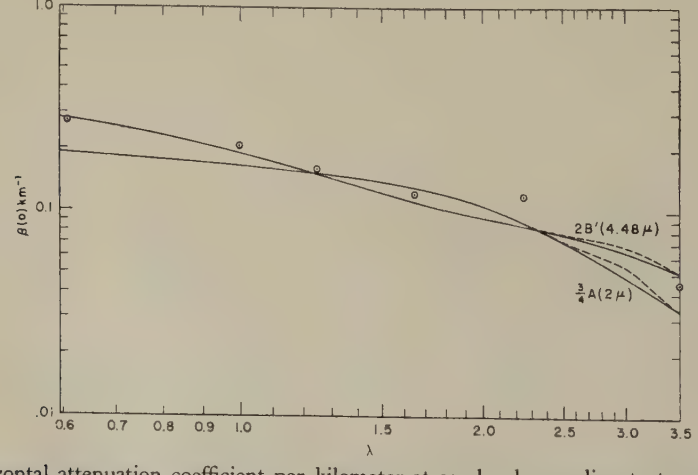

Figure 4. Horizontal attenuation coefficient per kilometer at sea level according to two models (full lines) and equivalent measurements by Gebbie *et al.* (circles). The effects of liquid-water absorption near 3 μ are indicated by the dashed portion.

(cf. Table 1). This is especially noticeable in the Model A aerosol, where the concentration does not diminish with the radius as fast as in the other models. The effect exists even in Models B and C, especially in the longer wavelengths.

The computed optical thickness can be compared with the dashed lines in Fig. 3, which represent data obtained from mountain stations (Aldrich and Hoover 1954). These have been reduced by means of the formula

$$\tau_M(h) = \text{antilog}_e[T(h)] - \tau_R(h) \quad . \quad . \quad . \quad (8)$$

where $T(h)$ is the reported monochromatic transmission, h is the station height above sea level and $\tau_R(h)$ is the corresponding Rayleigh atmosphere optical thickness. In Fig. 3 the label S.K. refers to Mt. St. Katherine in the Sinai Peninsula, about 29°N, 34°E, altitude 2·620 km; N.M. is for a station near Tyrone, New Mexico, 33°·5 N, 108°·3 W, altitude 2·438 km; and Table Mountain is in Southern California at 34°·3 N, 117°·7 W, altitude 2·286 km. The reported amounts of water vapour in the path of the sunlight, in mm of precipitable water, are shown in Fig. 3 in each case. Both in these observations as well as those shown in Fig. 2, there is evidence of a positive correlation between amounts of water vapour and the infra-red optical thickness of the haze, although the relation does not appear to be linear.

The fact that the plots in Fig. 3 are quite similar in general, despite the differences in geographical position of the observing stations, is a striking feature and must be significant in forming a concept of the hazy atmosphere. Some theoretical models, such as are proposed here, should be very useful in this respect. If we discount the ozone effect near λ 0·6 μ, these models, with minor variations in concentration, are quite capable of reproducing the observations. The deviations near λ 1·19 μ for the ' dry ' days, shown by the lowest dash curves in Fig. 3, need explanation. This particular feature of the haze attenuation could be attributed to some other constituent of desert air. However, we note from Fig. 1 that there is a strong water-vapour absorption band centered at λ 1·19 μ. The deviation from the theoretical curves at this wavelength also appears in the rest of the observations shown in Figs. 2 and 3. It is thus quite possible that the Smithsonian method of reduction has not entirely eliminated the water-vapour absorption here.

4. Comparison with more recent observations and discussion

Data pertaining to the detailed horizontal attenuation spectrum of infra-red radiation are not yet very abundant but more will undoubtedly be available in the near future. Among the best published data are those by Gebbie and co-authors (1953) based on observations along the eastern Scottish coast. The reported transmission per sea mile was converted to the attenuation coefficient per km (assuming that the Bouguer-Lambert exponential extinction law applies) and the result is shown by the dots in Fig. 4. Only the maximum transmission measured in a given water-vapour window was used. The full-line curves in Fig. 4 were fitted to the theoretical values for $\beta_R(0) + \text{const} \times \beta_M(0)$. The constant factor, 0·75 for Model A, and 2 for Model B′, simply modifies the total concentration of particles in the original model.

It will be seen that the curve 2B′ (4·48 μ) seems to give the best fit. This means that the observed transmission may be attributed to the presence of about 700 haze particles per cm^3 with a size distribution as in Model B′, Eq. (3), of which only 35 particles per cm^3 have a radius $0{\cdot}35\ \mu \leqq a \leqq 4{\cdot}48\ \mu$. This is only 5 per cent of the total concentration, and yet because of their large geometrical cross-section and their scattering properties, these few larger particles contribute 80 per cent or more of the attenuation. According to Gebbie *et al.*, the transmission in the visible and infra-red region decreases with increasing relative humidity. A higher humidity may be accompanied by both an increase in the total concentration and in the size range of haze particles. Our analysis shows that a simple increase in the overall concentration, without changing the size distribution and

the range, results in higher attenuation by a constant factor at all wavelengths. An extension of the size toward larger particles, without changing the distribution and concentration, results in larger overall attenuation, by a factor which increases with the wavelength. However, the work of Gebbie and co-authors (1953) suggests that an increased amount of water vapour is accompanied by an increased concentration in the smaller particles. This means that as the relative humidity increases, ever smaller atmospheric particles become effective as condensation nuclei. More observations of this type, made under varying conditions and compared with theoretical models, will help to establish the relation between humidity and the nature and concentration of natural haze.

An analysis of the scattering effect along the lines here presented affords a straightforward explanation of the observed window transmission in terms of haze models already proved in explaining the atmospheric transmission in the visible spectrum. The method further yields a good estimate of the concentration and size distribution of the particles, and these can be made to fit the observations by modifying well-defined parameters without having recourse to a changing index of refraction with wavelength (at least for wavelengths less than 2·18 μ).

With regard to other comparable measurements, the work of Arnulf and co-authors (1957) was concerned mainly with dense hazes and ground fogs involving particle concentrations and sizes much in excess of those considered here. The spectral detail shown in their plots is not sufficient in the range 1 μ to 2·25 μ for a comparison with other measurements and with theory. The evolution of the fogs and hazes in terms of the measured extinction, especially in what they call ' stable fogs,' suggests an increase in concentration without change in the distribution.

The measurements described by Taylor and Yates (1957) have not been reduced to obtain transmission or extinction coefficients. In the analysis of Elder and Strong (1953), which also deserves mention, the window transmission in a wide band is correlated with the peak transmission and the amount of precipitable water. The haze models here proposed explain the window transmission just as well, with the added advantage that the extinction by scattering on haze particles is separated from that caused by lines, and wings of distant strong lines, in the water vapour spectrum.

In the computed extinction curves shown in Figs. 2 to 4, the portion between λ 2 μ and 3·5 μ is not entirely correct because the liquid-water absorption around λ 3 μ was neglected. An indication of the effect of this absorption is shown in Fig. 4 by the dashed part of the curve, computed on the basis of Johnson and Terrel's (1955) values for a complex index of 1·33 (1 − 0·03 i). The effect represents a small correction in this case. The absorption becomes more important farther in the infra-red region and will be discussed in detail in a further paper on the basis of the appropriate complex index of refraction.

Appendix

The analytical approximation to the scattering cross-section, for the case $m \to 1$, is given by van de Hulst (1957) in the form

$$K(z) = 2 - \frac{4 \sin z}{z} + \frac{4 (1 - \cos z)}{z^2} \quad . \quad . \quad . \quad . \quad (9)$$

where $z = 2\alpha (m - 1)$ is the normalized size parameter and m is the real index of refraction. D. S. Saxon (unpublished work 1952) arrived at an identical expression based on somewhat different physical reasoning. The approximation has a wider range of validity than indicated by the original assumption, as shown by its author, by Deirmendjian (1957) and others. For the range $1 < m \leqq 1{\cdot}5$ the function (9) reproduces the position of the principal maxima given by the exact theory of G. Mie, but the value falls short of the exact, especially for relatively large values of m. Extensive comparison with the exact values computed by Penndorf (1956) shows that in the first maximum, occurring always at about $z = 4{\cdot}08$, the difference $K(\alpha, m) - K(z)$, where $K(\alpha, m)$ is the value given by

Mie's series, invariably amounts to $\frac{m-1}{m} K(z)$. This difference vanishes as z^{-1} for values $z > 4{\cdot}08$.

A modified formula such as

$$K'(z) = \left(1 + \frac{m-1}{m}\frac{4{\cdot}08}{z}\right) K(z), \quad z > 4{\cdot}08 \quad . \quad . \quad . \quad (10)$$

where $K(z)$ is given by function (9), yields very good results for intermediate values of z, with a difference not exceeding two per cent of the exact cross-section. Similarly, for smaller particles, the modified formula takes the form,

$$K'(z) = \left(1 + \frac{m-1}{m}\frac{z}{4{\cdot}08}\right) K(z), \quad 5(m-1) < z \leqq 4{\cdot}08 \quad . \quad (11)$$

When the integrals (6) are expressed in terms of the variable z and the functions (10) and (11) are used for the scattering cross-section, the integration can be carried out in closed form, provided the upper limit is finite. It was not necessary to use the analytical approximation here since the exact values for $m = 1{\cdot}33$ were already available in detail. However, it will be demonstrated in a further paper that the analytical approximation for complex index, also derived by van de Hulst, may be similarly modified with excellent accuracy. Its use then becomes important in estimating the infra-red extinction for particular values of the complex index without recourse to the cumbersome Mie series.

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551.509.313 : 551.511

The effect of stability on computed tendencies and vertical velocities

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(Manuscript received 8 January 1959)

SUMMARY

The two-parameter models used in numerical prediction of the contour height field usually assume a constant value of the hydrostatic stability, about $-0.04°C\ mb^{-1}$. The effect of varying this value over the field is investigated and it is shown that as the stability approaches zero the computed tendencies and the vertical velocities are increased.

1. INTRODUCTION

The models used in numerical weather prediction usually provide two relations between rate of change of the thickness of a layer and its mean vertical velocity. The first relation expresses the rate at which the vertical component of the vorticity of the wind speed is changing and is dynamical in origin; the second expresses the thermodynamical control of the rate of change of mean temperature or thickness. In the model described by Sawyer and Bushby (1953) these two equations are

$$\nabla^2 \frac{\partial h'}{\partial t} + J\left(h_m, \frac{g}{f}\nabla^2 h'\right) + J\left(h', \frac{g}{f}\nabla^2 h_m + f\right) = \frac{4\bar{\omega}_m(\nabla^2 h_m + f^2/g)}{p_0 - p_1} \quad . \quad (1)$$

$$\frac{\partial h'}{\partial t} = \frac{g}{f} J(h', h_m) - \frac{R}{g} A\, \Gamma\, \bar{\omega}_m \quad . \quad . \quad . \quad (2)$$

where the symbols are as defined in the original paper; it will suffice to recall here that h_m is the contour height at 600 mb, h' the 1,000-600 mb thickness, and $\bar{\omega}_m$ the vertical velocity at 600 mb. Γ is a measure of the static stability defined by

$$\Gamma = \frac{\partial T}{\partial p} - \frac{(\gamma - 1)\,T}{\gamma p} \quad . \quad . \quad . \quad . \quad (3)$$

and assumed to be constant in the vertical. The experimentally determined value of $-38°F/500$ mb was adopted when computations were made with this model. Recently Sawyer (1958) has shown that the effect of taking Γ to be zero in a local area where rain may be expected to fall can lead to appreciable modification in the thickness changes computed over a period of 12 hr.

If at any instant the values of h_m, h' are known, Eqs. (1) and (2) may be solved for $\partial h'/\partial t$ and $\bar{\omega}_m$; the values will depend upon the value chosen for Γ. The computation of thickness tendency and vertical velocity was carried out with five values of Γ, the standard value Γ_s of $-38°F/500$ mb and $3\Gamma_s/4$, $\Gamma_s/2$, $\Gamma_s/4$ and $\Gamma_s/32$. The initial fields of h_m and h' were the contour heights at 600 mb and the 1,000-600 mb thickness, each obtained by interpolation from the contour heights at 500 mb and the 1,000-500 mb thickness as constructed from the observations made upon eight occasions in 1957. The common factor in each of the eight occasions was the existence of a depression to the west of the British Isles which subsequently moved over the country.

2. The results

The computations gave the values of thickness tendency and vertical velocity at the nodes of a square mesh, there being 14 × 10 nodes; the nodal distance was about 100 mi, half that used in carrying out a time integration, the lesser distance being used in order to give more detail. The solution was carried out by eliminating ϖ_m between Eqs. (1) and (2) and solving the resulting Helmholtz equation for $\partial h'/\partial t$; the boundary values of $\partial h'/\partial t$ were taken to be those which actually occurred in the ensuing 12-hr period. The values of $\partial h'/\partial t$ so obtained were then inserted into Eq. (2) to yield ϖ_m.

As was to be expected, the results showed a well-defined region in which the thickness was increasing owing to the horizontal advection of warm air and another well-defined region in which the thickness was decreasing to the rear of the depression, although both changes were partially counterbalanced by cooling or warming due to vertical motion. From each of these regions an area showing large positive or negative tendencies was chosen as representative of the changes taking place. These areas, which were the same for each of the five different values of Γ but differed from occasion to occasion, contained between 6 and 12 nodal values to ensure that a freak value would not dominate the average which, however, should not be taken over too large an area; boundary points were avoided. The simple arithmetic average over these nodal points was computed. The vertical velocity fields were then examined for ordered values which were connected with these regions of maximum thickness tendency and the average value of the vertical velocity found over a common set of nodal points for each value of the stability, not necessarily the same set as used in finding the average tendency. For two sets of data the vertical velocities associated with negative tendency were small, and they have not been included in the diagram, because their values were so small as to make difficult distinction from the axis. The thickness tendencies must have been due almost entirely to advection.

In examining the computations it became clear that the fields of thickness tendency were smoothly distributed and there was no difficulty in selecting a representative area or in forming a representative mean. This was true of the field of vertical velocity except for the near-zero value of stability for which the vertical velocity sometimes became large with alternations of sign at neighbouring nodes, especially in the region of negative tendency. If ϖ_s, h_s' correspond to the value of Γ_s, and ϖ, h' to the value Γ then from Eq. (2)

$$\varpi = \frac{\Gamma_s}{\Gamma}\left\{\varpi_s + \frac{g}{RA\Gamma_s}\left(\frac{\partial h_s'}{\partial t} - \frac{\partial h'}{\partial t}\right)\right\} \quad . \quad . \quad . \quad (4)$$

where if $\partial h/\partial t$ is given in ft hr^{-1} and ϖ in mb hr^{-1}, $g/RA\Gamma_s = -0{\cdot}682$ mb ft^{-1}. The computed results show that $\partial h'/\partial t$ exceeds $\partial h_s/\partial t$ and their difference has the opposite sign to ϖ_s. Hence as Γ is increased towards zero, the factor Γ_s/Γ increases, while the term inside the bracket decreases, rather less rapidly; to ensure that ϖ is accurately computed the values of $\partial h_s'/\partial t$, $\partial h'/\partial t$ and ϖ_s must be computed more accurately as Γ approaches zero. Actually, the Helmholtz equation was solved with the same criteria for each value of Γ and the apparent computational errors in ϖ, evident from large values of opposite sign at adjacent points, may have arisen because the solution had not been carried out with sufficient accuracy. When the values of ϖ are in doubt, they have not been included in Fig. 1.

Fig. 1 shows graphically the variation of thickness tendency and of vertical velocity with change in stability. The regional means are plotted against stability, and each set of data referring to a particular day is distinguished by a symbol or by the line joining symbols. On the left are the regional means in the region where thickness is increasing and on the right where thickness is decreasing. The most striking feature is the regularity of the variations; in each case the form of the curve is similar. The magnitudes of the tendency and of the vertical velocity increase as the stability approaches zero, and the computed values of thickness tendency and ϖ have opposite signs, except in the two cases in which the values of ϖ have been omitted because they were small. The variation of tendency with stability is approximately linear in the range $(\Gamma_s/2, \Gamma_s)$ where the variation

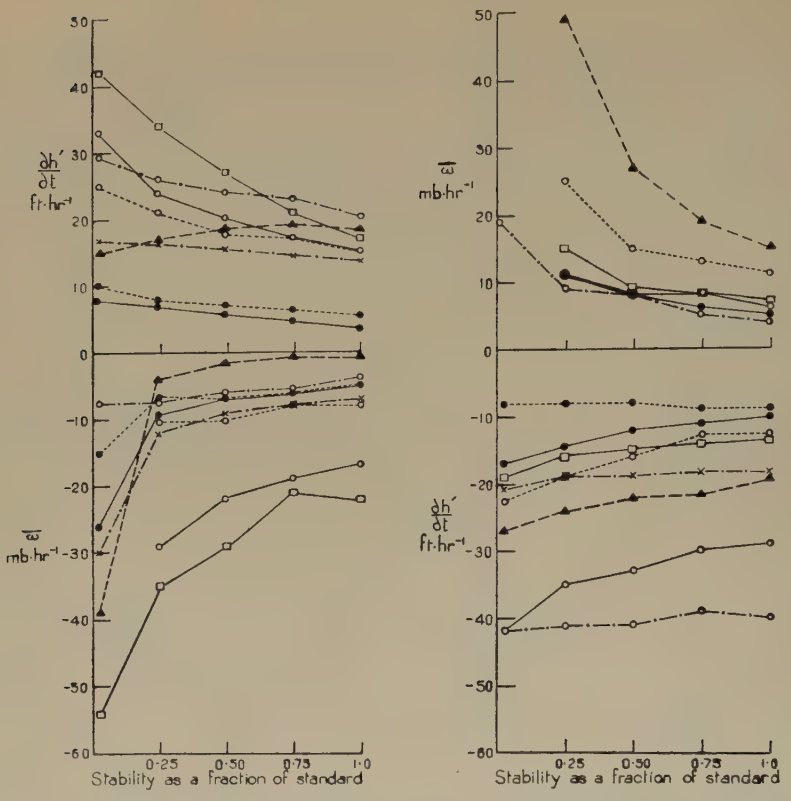
$\frac{\partial h'}{\partial t}$ fr·hr⁻¹
$\bar{\omega}$ mb·hr⁻¹
Stability as a fraction of standard
$\bar{\omega}$ mb·hr⁻¹
$\frac{\partial h'}{\partial t}$ fr·hr⁻¹
Stability as a fraction of standard

Figure 1. The variation of computed thickness tendency and vertical velocity with stability.

does not exceed about 10 per cent of the mean value. In the range $(\Gamma_s/32, \Gamma_s/2)$ th tendency increases more quickly than linearly. The variations in $\bar{\omega}$ follow the same patter approximately linear in the first range; but the variation is rapid as Γ approaches zer Since the standard value of Γ_s may be expected to be a good approximation to Γ ove large areas, values near zero and perhaps large gradients of Γ occurring only over sma areas, the local variations of Γ should play little role in determining the vertical velocit and thickness change over much of the area of the synoptic chart. In the small areas whe the value of Γ differs considerably from Γ_s the thickness tendency may well be augmente by 20 per cent or more and $\bar{\omega}$ by 50 per cent or more, so that the effect may be considerabl in areas in which Γ approaches zero, as in areas of rain.

Acknowledgment

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551.511.6 : 551.524.4 : 551.554

Estimation of surface stress and heat flux from profile data

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(Manuscript received 24 March 1959)

SUMMARY

The problem of estimating shearing stress from the wind profile in the lowest layers is discussed from the operational viewpoint. It is concluded that wind-profile data alone may never in practice be sufficient, even though a universal profile with the stress as a separable parameter may be established by future research. On the other hand the determination of both stress and heat flux, with adequate accuracy, may be possible from single values of temperature gradient and wind gradient observed simultaneously. A difficulty common to all indirect methods lies in interpreting the sampling variations.

1. INTRODUCTION

From the general meteorological viewpoint, the main purpose of measuring the profile of any quantity close to the ground is in the estimate of vertical flux which may then be obtained. With the continued development of numerical forecasting of the large-scale dynamical processes we may well envisage the need for routine assessment of such quantities as shearing stress and vertical heat flux as an operational requirement of the future. To be effective this would have to be achieved on a fairly widespread scale and be freed as far as possible from the idealisations as to site which have been a feature of most experimental programmes to date. The results would most suitably be framed, if it proves possible to do so, in terms solely of the bulk parameters of the situation (e.g. gradient or ' free-air ' wind, net incoming radiation, state of ground); but in this short note we shall be concerned with the problem of using, as reference data, measurements in the air at heights of the same order as normal screen level.

2. THE WIND PROFILE

It is now generally accepted that, when wind blows over rough ground under neutral conditions, the relationship between wind profile and shearing stress is given by

$$u = A \ln \frac{z}{z_0} \quad . \quad . \quad . \quad . \quad . \quad (1)$$

where $A = \frac{1}{k}\left(\frac{\tau}{\rho}\right)^{\frac{1}{2}}$. The notation is standard. Further, there is considerable convergence towards the view that a satisfactory generalization of Eq. (1) to small and perhaps to moderate degrees of thermal stratification exists in the form of the *log-linear* profile

$$u = A \ln \frac{z}{z_0} + B(z - z_0) \quad . \quad . \quad . \quad . \quad (2)$$

The inverse time B would characterize the degree of instability on the occasion in question, taking increasing positive or negative values for increasingly stable or unstable stratifications respectively. Monin and Obukhov (1954) have shown further that B should be given in terms of τ and the upward heat flux H by

$$B = -\alpha \frac{gH}{\tau c_p T} \quad . \quad . \quad . \quad . \quad . \quad (3)$$

where T is absolute temperature and α a universal constant.

Eq. (2) is not unique among possible first-order extensions of Eq. (1), but we shall not discuss here the basic implications and other merits of this particular choice. The equation will be used as an example of a tenable form for the universal wind profile, in order to discuss the usefulness and sufficiency of such a profile in the practical application defined in the opening paragraph.

Measurements of wind at three heights would, in an algebraic sense at least, be sufficient to determine all the parameters in the equation and hence τ and, subject to the Monin-Obukhov relation, H also. If more than three heights are used we may, by methods of regression, also obtain measures of the accuracy with which each of the parameters has been determined. On the face of it, an attractive and relatively simple field technique is promised for the routine synoptic assessment of both τ and H.

3. Operational test of surface stress

A test of the likely accuracy is possible by reference to the work of Rider (1954), whose measurements of wind profile are held to be among the best of their kind and were accompanied by simultaneous measurements of τ by a drag plate. The wind was averaged to the nearest cm per second over approximately half-hour periods at eight heights from 15 to 200 cm. These profiles have been re-analysed, firstly by determining best-fitting curves of the form

$$u = A \ln z + B z + C \qquad (4)$$

and secondly by fitting Eq. (2) as it stands with only two adjustable parameters A and B, z_0 being pre-assigned to the value 0·32 cm determined by Rider as that appropriate to his site. The value of A is then sufficient to determine τ, whence that of B provides H, while C provides z_0 in the first case. Table 1 provides fairly complete results of three profiles, one each under quasi-neutral, strong lapse, and moderate inversion conditions, chosen at random from within these groups. The standard errors of τ and B, as obtained by the process of curve fitting*, are included and the Richardson number at 1·5 m is quoted for reference.

TABLE 1. Parameters obtained by regression analysis

Run No.	$Ri_{1\cdot5}$	τ (dynes cm^{-2})			B (sec^{-1})		z_0 (cm)
		First method	Second method	Measured	First method	Second method	First method
11	− 0·014	6·57 ± 0·73	2·50 ± 0·15	3·73	− 0·46 ± 0·13	+ 0·42 ± 0·17	1·61
24	− 0·883	0·26 ± 0·04	0·20 ± 0·004	0·41	− 0·17 ± 0·10	− 0·11 ± 0·02	0·55
48	+ 0·108	0·34 ± 0·04	0·28 ± 0·004	0·19	+ 0·05 ± 0·03	+ 0·10 ± 0·015	0·48

Either method achieves a close fit to the wind observations and, as judged by this closeness of fit, a fairly precise value of τ emerges. But this appearance of accuracy is not supported by agreement between the profile-inferred and the measured values of τ, nor by mutual agreement between results obtained from the two methods of fitting. Thus the closeness of the curve fittings is deceptive, and provides no assurance of physical accuracy in the results inferred therefrom.

Looking next at the tabulations of B, and bearing in mind the deceptiveness of those of A, we must conclude that these results are almost worthless and that determination of heat flux from the wind profile is quite unreliable. This corresponds to the impracticability of identifying the value of the well-known Deacon β stability index from an individual wind profile (Deacon 1953, 1956).

In view of these preliminary findings, the bulk of Rider's data has been analysed only in respect of A, and without consideration of the standard error. The significant comparisons of τ are described hereunder, using suffices 1 and 2 for the two methods of reduction and zero for the drag-plate measurement.

Of 47 comparisons, 31 were in lapse and 15 in inversion conditions, one being indistinguishable. The geometric averages of $\frac{\tau_1}{\tau_0}$ were 1·16 and 0·80 in lapse and inversion respectively, those of $\frac{\tau_2}{\tau_0}$ were 0·82 and 0·85. Thus no major systematic error is in evidence. On an individual basis

* The method of least squares was employed and the standard errors obtained from Student's t distribution with 6 and 5 degrees of freedom respectively in the first and second methods, there being 8 levels of observation in each profile.

τ_1 and τ_0 differed by factors exceeding 1·5 on 23 occasions, exceeding 2 on 10 occasions and 3 on 6 occasions, these covering the whole range of stabilities. This cannot be regarded as an acceptable measure of agreement, though it might of course be held that there is no *a priori* assurance that the drag-plate values are the more nearly correct. However, with the second method the agreement was much better, τ_2 and τ_0 differing by factors exceeding 1·5, 2 and 3 on 11, 5 and 2 occasions respectively; the last two categories were confined to occasions of light winds, with τ_2 smaller than τ_0 in each instance. This strong convergence towards better agreement when proper restrictions are placed on z_0 does suggest that most of the disparity between Eq. (4) and the drag plate derives from errors in the first method. The implication is that surface stresses of acceptable accuracy can be achieved by wind-profile reductions of the type considered here only when z_0 is assigned in advance.

Such a requirement carries a grave operational disadvantage. It means that, even on carefully chosen and controlled sites, no individual profile can be interpreted until a large number have been measured, sufficient to calibrate the site. With present knowledge only neutral conditions can be used for calibration purposes and in sunny climates these conditions are rare, except at times of change which conflict with the steady-state requirement. Furthermore, the labour of calibration is increased many-fold over the more natural sites where z_0 must be expected to vary with stage of growth, with wind speed, and possibly with wind direction.

4. Discussion

It has been seen that, despite great experimental care and a most favourable site, from an individual profile without auxiliary information we are not justified in distinguishing between a number of fitted curves which carry very different physical implications. It seems that the same conclusion must hold for the temperature and humidity profiles in their parallel contexts. The essence of the difficulty is that whereas a reasonably accurate and representative measurement of the gradient can readily be made, the same is not true of the second derivative or ' shape ' of the curve. There is too much freedom in the fitting of the latter, and this may induce highly misleading results, at least unless a remote ' anchor ' to the curve is supplied (e.g. by assigning z_0), which may not be practicable in general. It is unlikely that the difficulty can be overcome by the development of more precise instruments, since sampling effects are probably the major cause and the more serious obstacle.

The importance of fixing z_0 is not to be interpreted as meaning necessarily that the roughness exerts a major influence at heights well away from the boundary. The writer shares the view that a profile for $\frac{\partial u}{\partial z}$ at these levels should exist which depends on τ and H but is insensitive to z_0 : that is to say, τ is more intimately associated with wind shear than with actual wind speed. This is one of the assumptions made in deriving the particular form of Eq. (2), but its validity has yet to be proven. The act of assigning z_0 before fitting the curve is a constraint to ensure against the adoption of a profile which implies impossible conditions at the boundary.

For the wind profile, failing z_0, the obvious source of auxiliary information is the simultaneous temperature profile. This opens too large a prospect for full discussion in a short article, but we must here note the strong possibility that determinations of both gradients (i.e. measurement of wind and temperature at two heights) may then prove sufficient for most practical needs. This involves, implicitly if not explicitly, the use of Ri as a criterion for the form of the relationship, and both stress and heat flux would then be separately determinate in terms of combinations of the two gradients. The most recent work along these lines is that of Deacon (1959) and, in the case of heat flux, of Taylor (1956). It must be emphasized that the true nature of such flux-gradient relationships can only be established from direct flux measurements, as in these last two studies, and not by inferential methods of obtaining the flux which are apt to beg significant aspects of the question.

Some indication of the individual accuracy of these latter results, or at least of their reproducibility which is really the quality under discussion here, may be obtained by comparing the heat flux as directly measured (Swinbank 1955; Taylor 1956) against the free and forced convection formulae expressed in terms of simple gradients (Priestley 1955; Taylor 1956). From 66 sets of fluctuation records at heights of 4 m and below, all in superadiabatic conditions, the measure of agreement is shown in Table 2. The tests described above for τ are also shown for comparison, now on a percentage basis.

TABLE 2. PERCENTAGE OF OCCASIONS WHERE FACTOR FALLS WITHIN STATED LIMITS

Flux considered	Two methods used	Factor			
		Within 1·5	More than 1·5	More than 2	More than 3
Momentum	Drag plate *vs* profile (first method)	51	49	21	13
Momentum	Drag plate *vs* profile (second method)	77	23	11	4
Heat	Eddy fluctuation *vs* gradient formulae	70	30	8	Nil

The comparison has limitations and should not be pressed too far. But in interpreting Table 2 it should be recalled that the first two rows refer to data averaged over 30-min periods as against 5-min periods for the data of the last row which, other things being equal, might therefore have been expected to lead to considerably more erratic results. That this does not happen supports the view that the less exacting basis of gradient measurements, compared with the detailed profile, offers the better promise of a practicable technique.

Discrepant factors such as those of Table 2 are due, apart from genuine error, to two distinct sampling effects. There is firstly the true variability of the flux, e.g. from one point of terrain to another ; and secondly that of the gradient (or profile) for a given flux or, in other words, the sampling variability of the gradient-flux (or profile-flux) relationship. The first is inherent, whereas the second is introduced by the indirect methods, and it is difficult to see how the two can then be separated. For this reason, as well as on the evidence above, the direct eddy-flux measurement itself must be considered as a possible operational tool of the future. Increasing automation should progressively remove most of the labour of analysis which has so far accompanied this method.

ACKNOWLEDGMENT

The writer is indebted to Mr. N. E. Bacon for the regression calculations on which much of this discussion is based.

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551.508.77

A rate-of-rainfall recorder

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(Manuscript received 8 June 1959)

SUMMARY

In certain atmospheric and geophysical studies it is desirable to have an accurate rate-of-rainfall recorder which can function for long periods unattended, which can respond within a few seconds to changes in rainfall and which can provide records of good time resolution. The instrument described in this paper was designed to meet these requirements.

1. THE INSTRUMENT

The principle of the instrument is simple. Rain is collected in the funnel of a conventional raingauge and allowed to accumulate in a nozzle at the base of the funnel to form large drops of constant size. These fall through a wire grid, alternate wires of which are connected together, and in so doing, complete a circuit which triggers a monostable multivibrator whose output is integrated to operate a pen-recorder with a signal proportional to the rate of rainfall.

Care has to be taken to ensure that the drops produced by the nozzle are of constant size. It was found that if the nozzle, which was constructed of brass, was heated to redness several times, an oxide surface layer was formed which was easily wetted. Above the tip, the bore must be constricted gradually to a throat which can be bridged readily by accumulating rainwater.

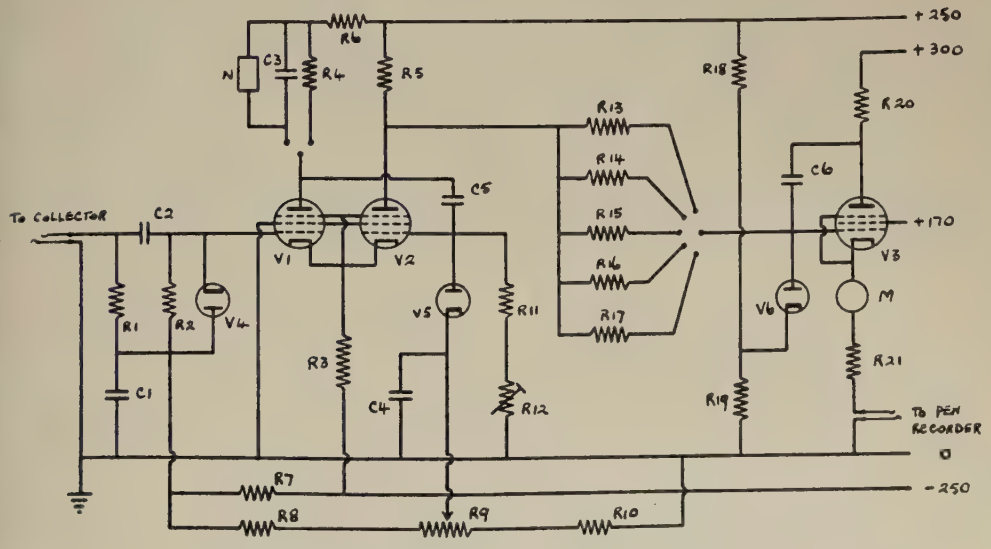

Figure 1. The electronic circuit for the rate-of-rainfall recorder.

C1	0·1 μf	R3	25 kΩ	R11	5 mΩ	R19	22 kΩ
C2	1,000 μμf	R4	2 kΩ	R12	5 mΩ	R20	60 kΩ
C3	0·25 μf	R5	30 kΩ	R13	3 mΩ	R21	9 kΩ
C4	0·1 μf	R6	27·2 kΩ	R14	2 mΩ	V1, V2, V3	EF91
C5	0·01 μf	R7	100 kΩ	R15	1 mΩ	V4, V5, V6	½ EB91
C6	4 μf	R8	4·7 kΩ	R16	500 kΩ	M	0–2 ma meter
R1	10 mΩ	R9	2·5 kΩ	R17	50 kΩ	N	counter (internal resistance 2 kΩ)
R2	1 mΩ	R10	20 kΩ	R18	220 kΩ		

Pen recorder resistance, 3 kΩ.

The wires of the grid are supported at a separation of about 4 mm on a polythene ring some 8 cm in diameter. To prevent corrosion by rain-borne matter carried from the somewhat polluted atmosphere in which the instrument usually operated, platinum had to be used for the wires.

The electronic circuit is shown in Fig. 1. It is that of a conventional cathode-coupled monostable multivibrator using two pentodes, the anode of one being directly coupled to a Miller integrator through one of several resistors providing time-constants of 1, 10, 20, 40 and 60 sec. The integrator acts as a cathode follower, driving a pen recorder over the range 0-2 ma. The total rainfall can be obtained by switching a counter into the anode of the other pentode. R9 provides zero adjustment by controlling the quiescent current through the multivibrator and R12 gives some adjustment of the pulse length and therefore of the calibration. The circuit as shown gives an output change of 1 ma for one pulse per second, which, with the rain collector used, corresponded to a rate of rainfall of 10 mm h^{-1}.

2. Performance

The high degree of negative feedback in the integrator and the fact that the shape of the pulses is unaffected by the repetition rate give the instrument a very good linearity. In all but the lightest rainfalls, negligible time lag is involved in the wetting of the funnel and the accumulation of water in the nozzle, so that the overall time constant differs little from the electrical time constant. Clean records are obtained with time constants down to 20 sec, although individual pulses can then be distinguished during low rates of rainfall. The instrument normally requires little attention other than the occasional changing of pen-recorder charts.

Fig. 2 shows the record obtained during a 2½ hr period of showery rain.

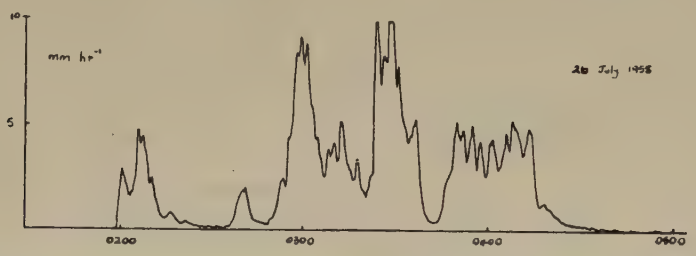

Figure 2. The rate of rainfall during a period of showery rain recorded with a time constant of 40 sec.

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Solar radiation and the duration of sunshine in Trinidad, West Indies

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(Manuscript received 22 June 1959)

SUMMARY

Records of daily solar radiation energy extending over a period of one year have been compared with readings of sunlight duration obtained with a Campbell-Stokes recorder. Energy records were made with a bimetallic antinograph. The regression equation obtained for the daily values in Trinidad, West Indies was $R = R_A(0{\cdot}20 + 0{\cdot}59\, n/N)$, with a correlation coefficient of 0·91

where R = radiation actually received

R_A = Angot's theoretical amounts of radiation which would reach the earth's surface in the absence of an atmosphere. These values are given by List (1951)

n = actual duration of sunlight

N = theoretical maximum duration of sunlight.

1. INTRODUCTION

In the estimation of plant water needs it is becoming recognized that the dominant factor which governs transpiration rate is the amount of energy received by the plant from solar radiation. Records of this quantity are not readily available for many areas of the world, particularly in tropical latitudes. Campbell-Stokes and the Marvin sunshine recorders are much more numerous and more likely to be found in under-developed areas than are instruments of the Ångström pyrheliometer type. It is therefore of interest to examine the correlation which may be presumed to exist between records of ' duration of bright sunshine ' as measured by such instruments and radiation intensity. It is believed that short-term results, such as those here presented, will prove of interest as an interim measure until records extending over a greater period can be provided.

Ångström (1924) originally suggested the relation

$$R = R_0\,(a' + (1{\cdot}00 - a)\, n/N) \qquad (1)$$

R_0 = radiation received on a completely cloud-free day

a' = mean fraction of the radiation received on a completely overcast day

Prescott (1940) has modified this and used the form

$$R = R_A\,(a + b\, n/N). \qquad (2)$$

where a, b = constants to be evaluated for any particular locality.

2. EXISTING RECORDS

Black, Bonython and Prescott (1954) appear to have made the most recent survey of stations throughout the world at which simultaneous records of solar radiation intensities and ' sunshine hour ' figures are available. For mean monthly values of these factors, they note that a can take values ranging from 0·15 to 0·40 and b from 0·27 to 0·61. The values obtained will depend on the type and amount of cloud cover. As an average of 32 stations, these workers give the regression equation

$$R = R_A\,(0{\cdot}23 + 0{\cdot}48\, n/N) \qquad (3)$$

3. Records from the West Indies

Until very recently, data of this kind have been available in the West Indies only from San Juan, Puerto Rico. Cowan (1959) in Jamaica obtained figures from which he deduced the regression equation

$$R = R_A (0{\cdot}31 + 0{\cdot}49\, n/N) \quad . \quad . \quad . \quad . \quad . \quad . \quad (4)$$

This relation is for daily figures, and not for mean monthly values as given by Black *et al.*

Through the courtesy of the West Indies Meteorological Service, a bimetallic actinograph of the British Meteorological Office type (Anon (1956), Blackwell (1953)) was obtained on loan. Since no standard or substandard instrument was available for calibration purposes in Trinidad, it had to be calibrated initially at Kew Observatory. Some difficulty was experienced in getting it out to Trinidad with the calibration unchanged, but it is believed that this was done successfully. It was set up on a flat roof at the Imperial College of Tropical Agriculture from which an almost uninterrupted view of the horizon was obtained. Daily records were made over a period of a little over a year, from August 1957 to August 1958. Total energy amounts were obtained by integration with a planimeter. This method, whilst laborious, was surprisingly efficient. Areas are obtained with a reproducibility of 0·1 to 0·2 per cent. The 'sunshine hours' instrument was a Campbell-Stokes recorder, installed within a few feet of the actinograph.

Fig. 1 shows the individual points plotted for each day of the period under investigation. The larger error possible at the lower end of the scale is especially noticeable, and agrees with results of other observers.

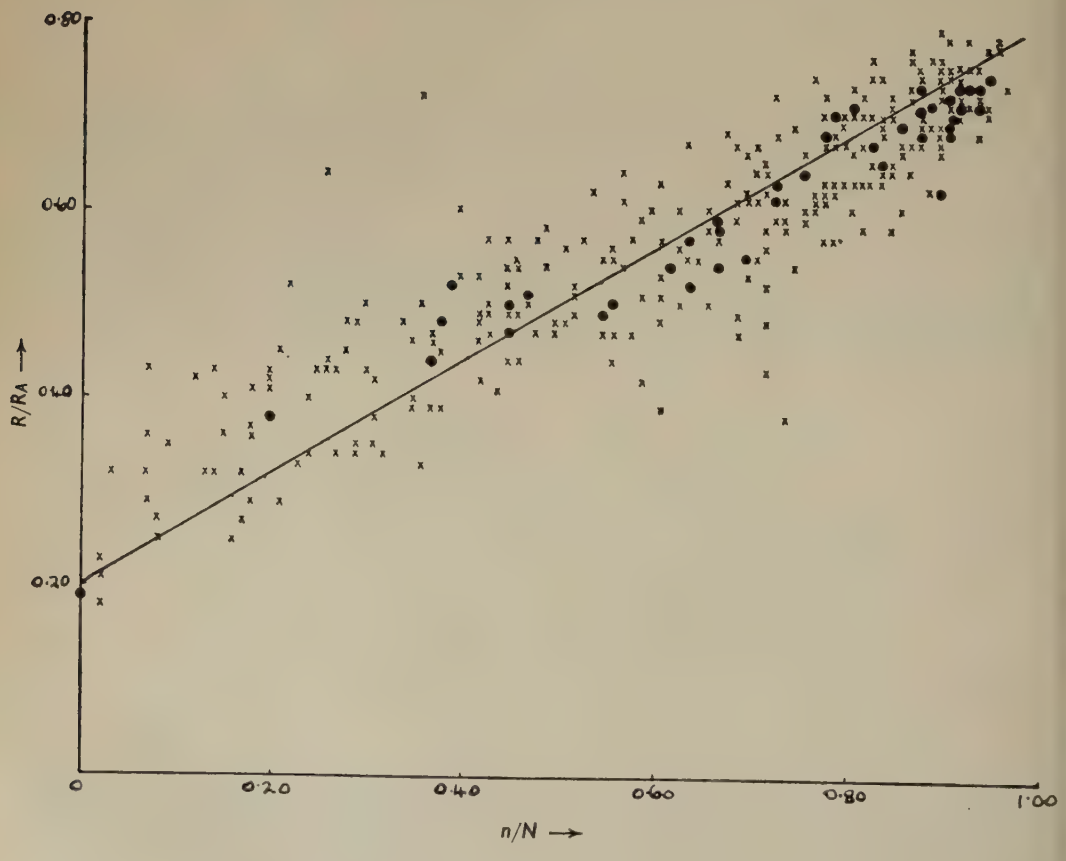

Figure 1. The relation between R/R_A and n/N for Trinidad, West Indies, using daily values.

A regression analysis was performed on the daily figures, giving the following equation

$$R = R_A\,(0{\cdot}20 + 0{\cdot}59\,n/N) \quad . \quad . \quad . \quad . \quad . \quad . \quad (5)$$

The correlation coefficient was 0·91. Since no detailed statistical procedures were to be attempted, an angular transformation was not performed on the raw data.

In the paper of Black *et al*, the nearest latitude to that of Trinidad (10° 38′N) is that of the group of stations comprising Batavia, Tjibodas and Bandung having latitudes ranging from 6°S to 7°S. For these stations the regression equation is given as

$$R = R_A\,(0{\cdot}29 + 0{\cdot}29\,n/N) \quad . \quad . \quad . \quad . \quad . \quad . \quad (6)$$

which is appreciably different from that of Trinidad given in the present paper.

Acknowledgments

I should like to express my thanks to Mr. W. E. Grinsted, who arranged the loan of the instrument, and to Mr. G. Day, who carried out the calibration at Kew Observatory.

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551.573 : 551.582 (267, 267.5, 267, 8)

Monthly charts of evaporation from the N. Indian Ocean (including the Red Sea and the Persian Gulf)

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(Communicated by H. Charnock. Manuscript received 14 May 1959)

Summary

Monthly values of the rate of evaporation from each 5-degree square of the N. Indian Ocean and from the Red Sea and the Persian Gulf have been obtained by using a modified form of Jacobs's formula and climatic data supplied by the Meteorological Office.

1. Introduction

Published information on evaporation from the N. Indian Ocean consists of the annual latitudinal values prepared by Wust (1936), charts for February and August by Albrecht (1949), the twelve-monthly charts by Budyko (1954) and the seasonal charts prepared by Venkateswaran (1956). For the Red Sea there are Vercelli's (1925) annual estimate, which is based on pan observations at coastal stations, and Neumann's (1952) estimates in the neighbourhood of 25°N 35°E; no attempt appears to have been made to compute evaporation from the Persian Gulf. In this paper 12 charts are presented showing the monthly distribution of evaporation from the N. Indian Ocean, monthly latitudinal profiles through the Red Sea and the annual variation of the rate of evaporation from the Persian Gulf. Ignoring the controversial question of the relative merits of the methods used in this and the above investigations to compute the rate of evaporation from the sea surface, it is thought that the results presented here may be an improvement on earlier attempts. For example, Budyko computed values for areas of size 5° lat. by 10° long. and used climatic data drawn from a variety of sources. Venkateswaran interpolated 5-degree square values for the wet-bulb depression from the charts contained in the United States Weather Bureau's publication *Atlas of Climatic Charts of the Ocean* and similar values for the air and sea-surface temperatures from *Monthly Meteorological Charts of the Indian Ocean* prepared by the Meteorological Office. The results presented in this paper are based on sets of observations consisting of simultaneous readings of the wet- and dry-bulb and sea-surface temperatures, all of which were made with standard instruments and at fixed hours.

2. Method and data

The equation $E = 0{\cdot}00587\ (e_s - e_a)\ w$ was used to compute the rate of evaporation, where E is the rate of evaporation expressed in gm cm^{-2} day^{-1}, e_s is the saturated vapour pressure (in millibars) corresponding to the sea-surface temperature and multiplied by 0·98 to correct for salinity, e_a is the atmospheric vapour pressure (in millibars) and w the wind speed in knots at the standard height of observation. This equation is the same as that derived by Jacobs (1951), but modified by a change in the value of the constant consistent with the use of British climatic data ; this modification is discussed by Privett (in press). The meteorological data used in this investigation were supplied by the Marine Division of the Meteorological Office, and consisted of surface observations recorded at six-hourly intervals aboard British ships during the period 1921–1950. The instrumental observations used were of the dry-bulb and wet-bulb temperatures in a screen on board and of the sea-surface temperature by bucket, and were made with a standard set of instruments by approved methods. Wind speed was estimated, and the mean speed in knots for each 5-degree square was obtained from the data used in the preparation of the wind roses shown in the Meteorological Office atlas. Monthly values of the rate of evaporation from each 5-degree square of the N. Indian Ocean were obtained by using the above equation and climatic data. No attempt was made to determine the areal distribution of evaporation from the Red Sea, for the density of meteorological observations is highest along the axis of the Sea, i.e., along the shipping route from Suez to the Strait of Bab-el-Mandeb. The observations were

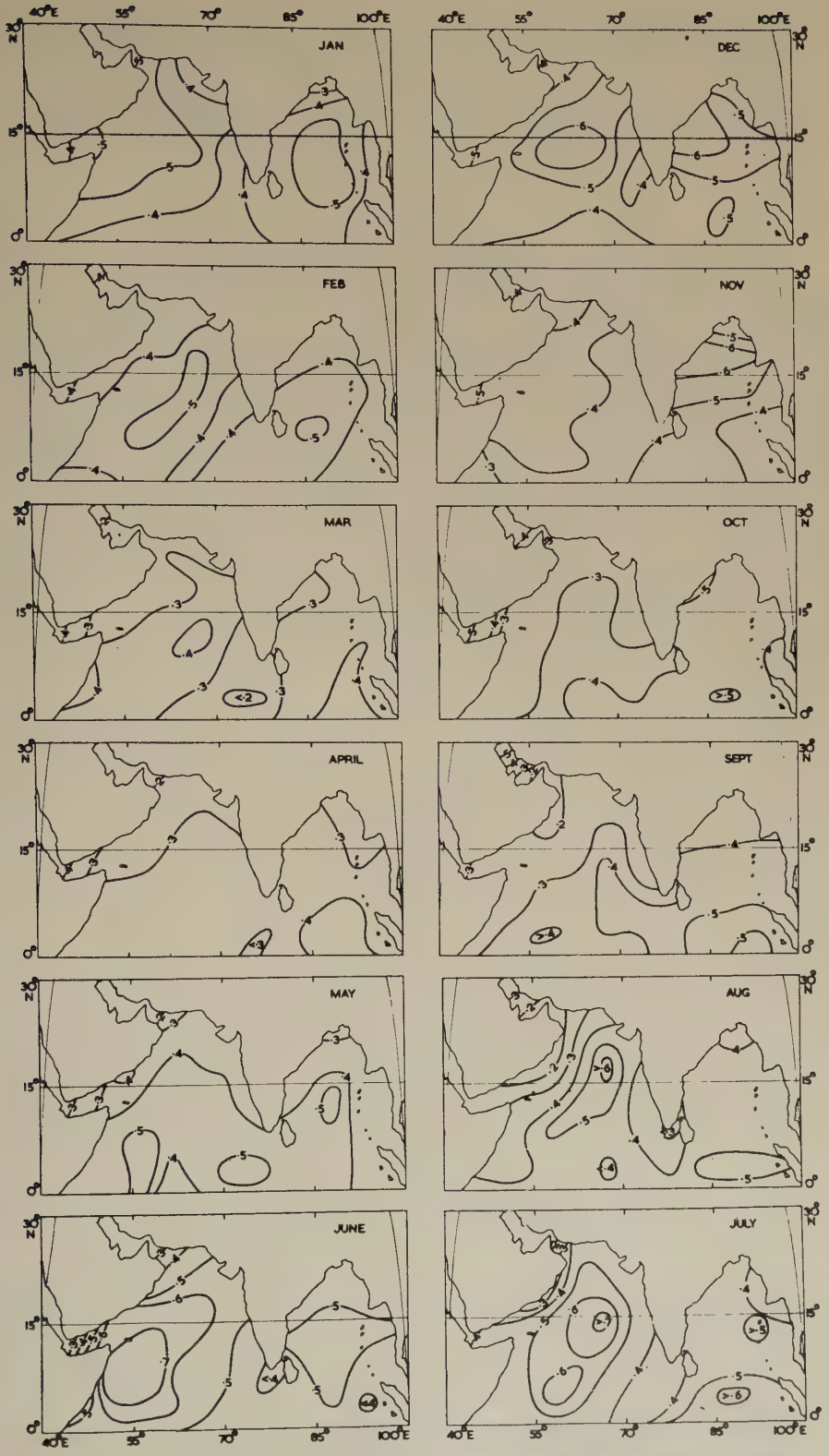

Figure 1. Monthly charts of evaporation (in gm cm^{-2} day^{-1}) from the N. Indian Ocean.

grouped into seven zones (the limits of these are shown in Fig. 3), and monthly values for the rate of evaporation from each of these zones were obtained. Estimates of evaporation were obtained for the areas to the north and south of 25°N in the Persian Gulf and these values were incorporated in the monthly charts shown in Fig. 1, but they should be treated with reserve as few meteorological observations were available. In view of this the diagram in Fig. 3 was prepared to show the annual variation for the Gulf when all the data were grouped together.

3. Discussion

(*a*) *The North Indian Ocean*

The variation through the year of evaporation from the N. Indian Ocean is shown in the monthly charts in Fig. 1. Over most of the oceans of the world it has been found that evaporation is highest in winter, least in summer, and greater in autumn than in spring, and that the areas of highest evaporation are located in the tropical trade-wind belts and over the warm ocean currents of middle latitudes in winter. This regime is found in the S. Indian Ocean, particularly to the south of 20°S as shown by Privett (in press), but to the north of the Equator a different rhythm is observed. During the year there are two periods of maximum evaporation, and the timing of these periods almost coincides with the maximum development of the NE and SW Monsoons.

Evaporation is low in the months March-April and September-October. During the latter period it is nearly everywhere lower in the Arabian Sea than in the Bay of Bengal. An absolute minimum of 0·18 cm day^{-1} is found in September off the coast of Oman in the north-west of the Arabian Sea. By November a marked increase in the rate of evaporation has taken place over nearly all the N. Indian Ocean. The largest increases are found in the Bay of Bengal, where an absolute maximum of 0·66 cm day^{-1} is found on the western side of the Bay and off the coast of Oman, where the rate has increased by 0·26 cm day^{-1}. This increase between the October and November values is clearly related to the emergence and establishment of the north-easterly air stream, with its relatively high wind speeds and low moisture content. In December evaporation reaches a maximum; the highest increase on the November values is found in the Arabian Sea where a maximum value of 0·65 cm day^{-1} has been obtained. Throughout January and February there is a gradual decrease in the rate of evaporation despite the continued presence of the NE Monsoon, which is said to reach its maximum development in these months. Evaporation again reaches a minimum in all areas in either March or April; an absolute minimum of 0·18 cm day^{-1} for this period is found in the Gulf of Oman in April.

This low rate of evaporation in March-April can be associated with the period of indeterminate winds which occur between the ' retreat ' of the NE Monsoon and the ' emergence ' of the SW Monsoon. From April to May there is an increase in evaporation particularly in the southern part of the ocean, and the chart for May shows a ridge of high evaporation extending northwards across the Equator off the coast of Africa; this is perhaps an indication of a transfer of air from the Southern Hemisphere. Concurrent with the development of the SW Monsoon the rate of evaporation increases everywhere and reaches a maximum during June and July; an absolute maximum of 0·78 cm day^{-1} is found off the coast of Africa in June, but for the whole ocean a maximum is reached in July. By August the rate of evaporation is beginning to decrease, although there are isolated areas where an increase from the July value is to be observed, such as in the NE Arabian Sea. This decrease in evaporation continues throughout September and into October, when the SW Monsoon suddenly collapses and a period of indeterminate winds again prevails before the emergence of the NE Monsoon.

In addition to this semi-annual variation the pattern of the isopleths in these monthly charts shows some other interesting features. Firstly, in nearly every month there are areas of high evaporation centred over the Bay of Bengal and the Arabian Sea, and these two areas together with the intervening trough are aligned SW to NE. This pattern is most pronounced during the Monsoons and is presumably to be associated with the low moisture content and the high wind speeds of the NE and SW Monsoons respectively. Secondly, the region of high evaporation ' moves ' in the direction of the airstream. This is particularly noticeable in the Arabian Sea, where between May and August the region gradually ' moves ' from SW to NE. Thirdly, the annual distribution indicates that evaporation is slightly higher on the average from the Bay of Bengal than from the Arabian Sea. This difference is greatest during the period August-November, but during the period May-July this pattern is reversed, evaporation being much higher from the Arabian Sea than from the Bay of Bengal.

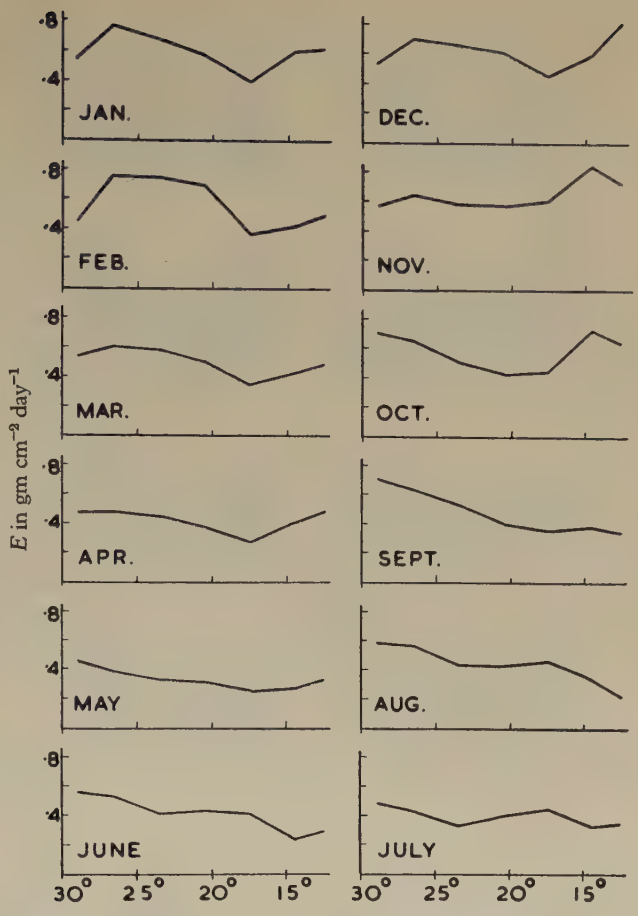

Figure 2. Monthly latitudinal profiles of evaporation (in gm cm^{-2} day^{-1}) from the Red Sea.

(b) The Red Sea and the Persian Gulf

The profiles in Fig. 2 indicate that from October to May evaporation tends to be higher towards the extremities of the Red Sea; the intervening low values are most pronounced in the zone 16°-19°N during the period December-April. During the remainder of the year there is generally a gradual decrease from north to south. This annual variation appears to be related to a reversal of the wind system. In the north the winds are predominantly NW throughout the year, but in the south these winds only occur from June to September, and for the remainder of the year a SE wind prevails. Between these two belts there is an area of light variable winds, which coincides with the region of low evaporation. Although evaporation for the whole Sea is at a maximum in November and at a minimum in May, the annual distributions shown in Fig. 3 indicate that this does not apply in every zone. For example, in the zones 25°-28°N and 22°-25°N maximum evaporation occurs in January and February respectively. An absolute maximum value of 0·85 cm day^{-1} occurs in November in the zone 13°-16°N and an absolute minimum of 0·24 cm day^{-1} in June in the same zone.

If these results are representative of the average rate of evaporation from each zone the mean annual evaporation from the Red Sea amounts to 183 cm, which is considerably lower than the estimate of 310 cm computed by Vercelli (1925) from pan observations at coastal stations. By interpolation of these results the mean annual evaporation at 25°N is 204 cm, which agrees favourably with the estimate of 210 cm for the same latitude reported by Neumann (1952) which was obtained by both the energy budget and aerodynamic techniques.

In the Persian Gulf the mean annual evaporation is 144 cm. The diagram in Fig. 3 shows that evaporation is at a maximum in December and at a minimum in May. These figures refer

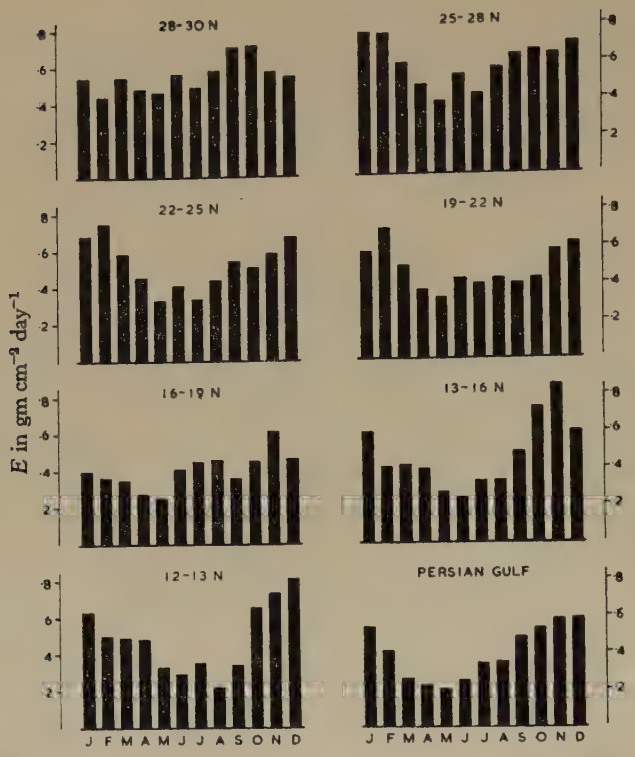

Figure 3. The annual variation in evaporation (in gm cm^{-2} day^{-1}) from seven latitudinal zones of the Red Sea, and from the Persian Gulf.

to mean conditions for the whole of the Gulf, but if the region is divided into northern and southern sections it would appear that from June to February there is an increase in evaporation from south to north ; there is some indication of this in the charts in Fig. 1.

Acknowledgments

I am indebted to the Director-General of the Meteorological Office and to the Superintendent of the Marine Division of the Meteorological Office for the loan of the data used in this investigation.

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551.508.21

On the influence of thickness and temperature on the 'cut-off' characteristics of glass filters

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(Manuscript received 29 June 1959)

SUMMARY

Schott glass filters OG1, RG2 and RG8 are investigated with respect to the dependence of the lower sharp wavelength 'cut-off' on filter thickness and temperature. Typical values of the main cut-off characteristics are presented conveniently in tabular form.

1. INTRODUCTION

The sharp cut-off Schott filters OG1, RG2 and RG8 are now widely used in meteorological radiation research as a means of separating a well-defined portion of the solar spectrum from one of shorter wavelength. The general transmission characteristics of this series of filters are illustrated in Fig. 1 which is an original record representative of an examination of the RG2 and RG8 type. (N.B. the wavelength scale requires correction by $-$ 3·5 mμ which has been applied in the results that follow). The transmittance, τ_λ, of a specific filter is related to the transmission, p_λ, and the reflectance (from two surfaces) through equation

$$\tau_\lambda = (1 - r_\lambda) \cdot p_\lambda \quad . \quad . \quad . \quad . \quad (1)$$

For most purposes, the reflectance, r_λ, can be regarded as independent of wavelength. We then have

$$p_\lambda = \frac{\tau_\lambda}{1 - r} \quad . \quad . \quad . \quad . \quad . \quad (2)$$

2. EXAMINATION OF RESULTS

The filters discussed here are characterized by (a) a spectral region of very high transmission – practically 100 per cent for $\lambda > \lambda_1$, (b) a region of almost total opaqueness for $\lambda < \lambda_2$, and (c) an intermediate region of cut-off $\lambda_1 > \lambda > \lambda_2$ within which the transmission increases rapidly with increasing wavelength. If we define the centre of cut-off as the wavelength λ_m for which the transmission is 50 per cent and λ_1 and λ_2 as the wavelengths for which the transmission is respectively 95 and 5 per cent, we may present the principal optical characteristics of the three filters through the following data :

	λ_m	λ_1	λ_2	$\lambda_1 - \lambda_2$	Thickness l	Temperature t
OG1	532	552	522	30	2 mm	20°C
RG2	627	647	615	32	2 mm	20°C
RG8	695	720	682	38	2 mm	20°C

The filters were investigated with regard to the dependence of the transmission on wavelength, filter thickness and filter temperature, with the aid of a Perkin-Elmer ratio-recording double monochromator of very high resolution and capable of a scanning speed of less than one minute for the whole visible region of the spectrum (Spectracord model 4000). Details of the instrumentation and technique employed have been given elsewhere (Ångström and Drummond 1959).

The effective bandwidths used were 1 and 10 mμ, respectively, in the visible and infra-red regions.

Two important relationships were revealed at an early stage of the investigation. They may be summarized as follows :

(1) The value of λ_m is a linear function of temperature, the centre of cut-off being displaced to longer wavelengths with increase in filter temperature. The shift in λ_m for a 10°C change amounts to : OG1, 1·2; RG2, 1·4; and RG8, 1·8 mμ. This shift is independent of thickness l, for values of l between 1 and 5 mm.

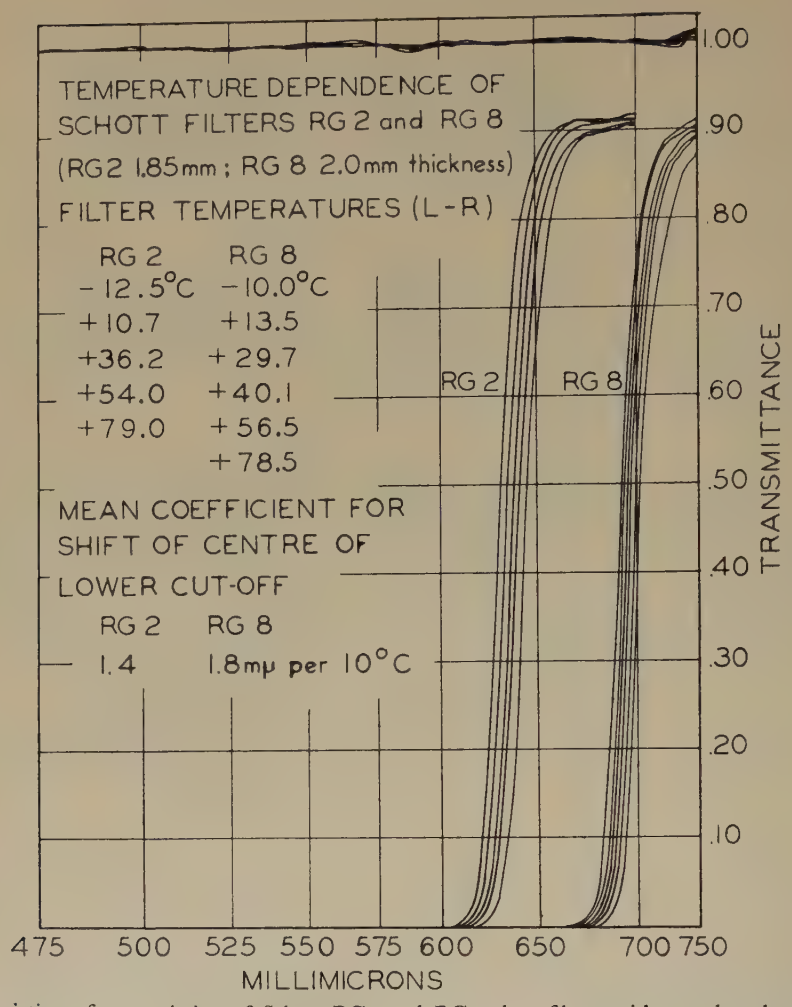

Figure 1. Variation of transmission of Schott RG2 and RG8 glass filters with wavelength at diffe temperatures.

(2) When values of log (− log p_λ) within the cut-off region are plotted against wavelen we find a linear relationship with a slope practically independent of temperature, t.

Combining these results, we arrive at the following general relationship :

$$p_\lambda = e^{-kl\,\phi(t)\,f(\lambda)} \quad . \quad . \quad . \quad . \quad .$$

where k is a constant (dependent on the value of λ which we choose as the starting poi The functions $\phi(t)$ and $f(\lambda)$ are

	$\phi(t)$	$f(\lambda)$
OG1	$e^{+0\cdot0168t}$	$e^{-0\cdot14\lambda}$
RG2	$e^{+0\cdot0182t}$	$e^{-0\cdot13\lambda}$
RG8	$e^{+0\cdot0194t}$	$e^{-0\cdot11\lambda}$

It may be derived from Eq. (3), and confirmed through experiments, that the shift in cut-off centre, λ_m, with thickness, l, of the filter relative to the position corresponding t thickness of 1 mm can be expressed by

$$\lambda_m = \lambda_m{}^1 + \frac{\ln l}{\gamma} \quad . \quad . \quad . \quad . \quad .$$

where γ has the following values : OG1, 0·14; RG2, 0·13; and RG8, 0·11.

For the transmission as a function of the wavelength difference, $\lambda - \lambda_m$, viz. the transmission around the centre of cut-off, we get

$$\left.\begin{array}{ll} \text{OG1} & p_\lambda = e^{-0\cdot69}e^{-0\cdot14(\lambda-\lambda_m)} \\ \text{RG2} & = e^{-0\cdot69}e^{-0\cdot13(\lambda-\lambda_m)} \\ \text{RG8} & = e^{-0\cdot69}e^{-0\cdot11(\lambda-\lambda_m)} \end{array}\right\} \quad . \quad . \quad . \quad (5)$$

From these relationships, derived on the basis of measurements for filter thicknesses of 1 to 4 mm and for filter temperatures between − 20 and about + 90°C, Tables 1 and 2 have been computed.

The filters examined are from melts specially prepared by Schott at the request of the Radiation Commission of I.U.G.G.

TABLE 1. POSITION OF CENTRE OF CUT-OFF (MILLIMICRONS)

l	1 mm			2 mm			3 mm			4 mm		
t(°C)	OG1	RG2	RG8	OG1	RG2	RG8	OG1	RG2	RG8	OG1	RG2	RG8
− 20	522·0	616·5	682·5	527·0	621·0	688·5	530·0	624·0	692·5	532·5	626·5	695·5
− 10	23·5	18·0	84·0	28·5	22·5	90·5	31·5	25·5	94·5	34·0	28·0	97·5
± 0	24·5	19·5	86·0	29·5	24·0	92·0	32·5	27·0	96·0	35·5	29·5	699·0
+ 10	26·0	21·0	88·0	31·0	25·5	94·0	33·5	28·5	98·0	36·5	31·0	701·0
+ 20	27·0	22·5	89·5	32·0	27·0	95·5	35·0	30·0	699·5	37·5	32·5	02·5
+ 30	28·0	23·5	91·0	33·0	28·0	97·0	36·0	31·5	701·0	38·5	34·0	04·0
+ 40	529·5	625·0	693·0	534·5	629·5	699·0	537·5	632·5	703·0	540·0	635·5	706·0

TABLE 2. TRANSMISSION IN REGION OF CUT-OFF

	$\lambda_m - 15$	$\lambda_m - 10$	$\lambda_m - 5$	$\lambda_m \pm 0$	$\lambda_m + 5$	$\lambda_m + 10$	$\lambda_m + 15$	$\lambda_m + 20$
OG1	0·005	0·06	0·25	0·50	0·71	0·84	0·92	0·96
RG2	0·01	0·08	0·27	0·50	0·70	0·83	0·905	0·95
RG8	0·03	0·13	0·30	0·50	0·67	0·80	0·875	0·925

The manner in which such filters are generally employed in meteorological research implies that the required short-wave solar radiation is computed as a difference between the total measured intensity (without any filter) and that transmitted by the particular filter. On the basis of our experiments and their treatment, we find that a 10°C rise in filter temperature results in an increase in the computed short-wave radiation of 0·8 and 1·3 per cent in the case of the RG2 and OG1 filters, respectively. An increase in filter thickness of 1 mm (1 mm $< l <$ 3 mm) produces increases of 2·0 and 3·7 per cent, respectively, for these filters.

When intensities within relatively narrow spectral bands are computed as differences between energies transmitted by different filters, considerably greater errors (as percentages) may arise unless proper consideration is paid to filter temperature and thickness.

Finally, it may be emphasized that the results reported above, refer specifically to certain glass melts furnished by Schott. The filters investigated here differ somewhat from those supplied earlier with similar designation and examined by Hinzpeter, Schulmann, Linke (1953) and Kimball and Hand (1958). There are reasons to believe, however, that the manufacturer will be able to closely reproduce, in future, the filter characteristics given in this study.

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551.521.11 (41/42)

An analysis of sunshine values in the British Isles for the period 1909-1958

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(Manuscript received 27 May 1959)

The Tables 1-5 which follow conclude my investigations into bright sunshine in the British Isles, which now cover a 50-yr. period. The values are prepared in the manner adopted previously (Hancock 1935, 1951). Maximum and minimum values are shown in the tables in bold face and italics respectively; where no maximum or minimum appears this occurred in the earlier period, 1909-48.

Table 5 shows the years 1909-1958 arranged according to the yearly sunshine values (in hours per day) for the British Isles as a whole.

TABLE 1. BRIGHT SUNSHINE AS A PERCENTAGE OF THE 1909-1933 MEAN

ENGLAND AND WALES

	Jan.	Feb.	Mar.	Apr.	May	Jun.	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1949	113	**161**	104	115	118	125	128	119	107	96	95	110	117
1950	78	101	111	104	93	122	102	99	78	93	87	98	97
1951	93	98	79	120	89	115	109	90	78	105	95	120	99
1952	**154**	118	81	108	106	97	100	102	85	95	115	124	107
1953	86	93	121	116	109	76	103	116	105	86	76	77	97
1949-1953	105	114	99	113	103	107	108	105	91	95	94	106	103
1954	116	87	87	122	83	70	73	74	110	78	89	108	91
1955	74	122	138	110	106	84	148	121	111	116	89	108	111
1956	116	102	125	110	127	79	92	90	72	110	96	53	99
1957	109	122	85	106	112	**144**	81	87	79	84	104	128	103
1958	114	83	99	101	94	72	98	86	94	94	76	87	91
1954-1958	106	103	107	110	104	90	98	92	93	96	91	97	99
Max. hr/day	2·31	3·79	5·79	7·54	8·54	9·51	9·45	8·65	6·49	4·30	2·91	1·89	4·63
Year	1952	1949	1929	1914	1909	1957	1911	1947	1911	1921	1920	1946	1949
Min. hr/day	0·95	1·17	2·17	3·03	3·69	4·39	3·63	2·88	3·11	2·19	1·19	0·75	3·35
Year	1917	1947	1936	1920	1932	1909	1944	1912	1945	1915	1934	1912	1912
Mean hr/day 1909-1958	1·51	2·29	3·64	5·15	6·22	6·54	5·73	5·40	4·45	3·16	1·95	1·30	3·94

TABLE 2. BRIGHT SUNSHINE AS A PERCENTAGE OF THE 1909-1933 MEAN

SCOTLAND

	Jan.	Feb.	Mar.	Apr.	May	Jun.	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1949	90	134	112	93	112	140	101	101	114	86	88	89	105
1950	106	136	99	95	106	98	112	115	85	112	102	97	105
1951	115	96	82	108	129	110	88	90	103	103	79	129	103
1952	**131**	112	102	111	102	81	112	89	86	113	120	107	106
1953	106	108	122	104	97	92	72	102	92	87	77	109	97
1949-1953	110	117	103	102	109	104	97	99	96	100	93	106	103
1954	86	89	106	123	86	81	77	75	98	80	84	78	89
1955	110	147	118	120	128	105	**168**	132	108	99	98	127	122
1956	120	94	122	127	103	104	107	92	79	117	84	67	101
1957	121	142	63	114	109	136	69	92	97	86	74	137	103
1958	116	83	97	94	108	72	109	90	107	100	92	130	100
1954-1958	111	111	101	116	107	100	106	96	98	96	86	108	103
Max. hr/day	1·52	3·29	5·26	6·89	7·77	8·57	7·91	8·22	4·96	3·53	2·40	1·44	4·14
Year	1931, '33, & '52	1930	1929	1942	1946	1940	1955	1947	1933	1922	1925	1935	1955
Min. hr/day	0·65	1·32	1·86	3·21	3·92	3·01	3·13	2·20	2·91	1·61	1·06	0·32	2·83
Year	1944	1923	1928	1937	1944	1912	1931	1912	1930	1935	1939	1912	1912
Mean hr/day 1909-1958	1·16	2·13	3·27	4·70	5·68	5·80	4·69	4·26	3·66	2·56	1·57	0·94	3·37

TABLE 3. BRIGHT SUNSHINE AS A PERCENTAGE OF THE 1909-1933 MEAN

IRELAND

	Jan.	Feb.	Mar.	Apr.	May	Jun.	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1949	96	111	99	86	112	142	92	101	120	89	124	94	106
1950	99	94	101	88	123	105	107	117	79	100	134	122	106
1951	106	132	87	113	117	129	96	98	76	87	88	112	103
1952	128	107	94	99	112	96	81	98	107	120	91	120	104
1953	94	101	120	111	105	98	102	96	81	101	86	78	98
1949-1953	105	109	100	99	114	114	96	102	93	99	105	105	103
1954	95	82	80	105	95	80	70	104	108	71	104	78	89
1955	109	**144**	144	113	113	*58*	162	114	115	122	82	96	114
1956	113	126	92	105	114	105	90	109	*60*	113	81	78	99
1957	152	156	*61*	115	128	**156**	73	88	74	81	99	124	109
1958	142	88	76	84	104	71	97	90	88	93	75	106	93
1954-1958	122	119	91	104	111	94	98	101	89	96	88	96	101
Max. hr/day	2·36	3·32	6·01	7·75	7·80	8·94	7·97	8·29	5·86	3·58	3·04	1·86	4·55
Year	1933	1955	1929	1938	1935	1957	1911	1947	1933	1939	1909	1938	1911
Min. hr/day	1·00	1·32	2·22	2·56	3·95	3·34	2·99	3·03	2·51	1·73	1·09	0·90	3·25
Year	1915	1926, '29	1957	1937	1925	1955	1944	1932	1956	1916	1917	1940, '47	1920
Mean hr/day 1909-1958	1·54	2·33	3·54	5·17	6·05	5·82	4·88	4·64	3·92	2·82	1·88	1·29	3·66

TABLE 4. BRIGHT SUNSHINE AS A PERCENTAGE OF THE 1909-1933 MEAN

BRITISH ISLES

	Jan.	Feb.	Mar.	Apr.	May	Jun.	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
1949	102	**142**	105	101	115	132	111	110	112	92	102	101	110
1950	89	108	105	97	104	112	104	108	80	99	105	105	101
1951	101	108	82	115	106	117	101	92	83	101	90	119	101
1952	140	114	89	106	107	93	97	98	91	106	109	120	106
1953	88	99	121	111	105	86	95	108	96	89	79	82	97
1949-1953	104	114	100	106	107	108	102	103	92	97	97	105	103
1954	103	86	90	117	87	75	72	82	107	77	92	92	90
1955	90	135	121	113	112	82	**155**	122	111	114	89	108	113
1956	115	108	115	112	118	91	94	95	***70***	112	89	63	99
1957	122	138	73	111	116	144	77	88	82	84	96	128	105
1958	91	***64***	93	94	99	72	99	83	96	95	79	100	89
1954-1958	104	106	98	109	106	91	99	94	93	96	89	98	99
Max. hr/day	2·01	3·20	5·72	7·03	7·77	8·98	8·26	8·44	5·67	3·73	2·74	1·66	4·39
Year	1933	1949	1929	1914	1948	1940	1955	1947	1911	1919	1925	1946	1911
Min. hr/day	1·00	1·43	2·18	3·07	4·34	4·30	3·36	2·76	3·00	2·19	1·23	0·71	3·23
Year	1914	1958	1936	1937	1932	1923	1944	1912	1956	1916	1912	1912	1912
Mean hr/day 1909-1958	1·43	2·26	3·52	5·04	6·04	6·16	5·24	4·91	4·11	2·92	1·84	1·21	3·72

TABLE 5. YEARS IN ORDER OF AMOUNT OF BRIGHT SUNSHINE FOR BRITISH ISLES, HR/DAY

1911	4·39	1948	3·84	1956	3·73	1936	3·49
1955	4·27	1928	3·83	1910	3·72	1941	3·47
1949	4·19	1925	3·82	1953	3·70	1924	3·46
1921	4·12	1934	3·82	1942	3·68	1931	3·46
1929	4·12	1940	3·82	1930	3·67	1913	3·45
1933	4·00	1950	3·81	1946	3·65	1932	3·43
1914	3·94	1943	3·80	1926	3·64	1916	3·42
1919	3·94	1952	3·80	1923	3·62	1944	3·42
1915	3·91	1951	3·79	1938	3·61	1937	3·38
1935	3·88	1922	3·78	1927	3·60	1958	3·33
1909	3·86	1917	3·75	1945	3·54	1954	3·31
1957	3·86	1939	3·75	1920	3·49	1912	3·23
1918	3·85	1947	3·73				

ACKNOWLEDGMENT

The author wishes to thank the Director-General of the Meteorological Office for his kindness in supplying the data for Eire as these are no longer published in the *Monthly Weather Report* (Meteorological Office).

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551.501.81 : 621.396.9

Radar echoes from atmospheric inhomogeneities

By W. G. Harper

Mr. R. F. Jones, in his recent contribution to this *Journal* (*Q.J.*, **84**, p. 437, 1958), has done ar meteorologists a valuable service. It is a number of years since ' angel '-type radar echoes re first attributed to meteorological causes, yet little theoretical support for this contention until now been brought forward. A simple solution on these lines cannot be expected, but el impelled to mention what I think are serious difficulties in accepting any form of side-lobe ution. Mr. Jones has overcome a number of awkward points in propounding his specialized n of reflecting surface, yet observational difficulties arise :

(a) If a side-lobe mechanism is invoked, a necessary corollary must be that a complete e-lobe pattern showing fixed relative intensity variations is displayed at the precise range ach ' angel '-type echo. This effect has not been observed.

While agreeing that some energy must leak out at substantial angles to the main lobe of iation I would contend that the pattern of the first few side-lobes is reasonably well known all high-power 10 cm radars. It is seen in echoes from aircraft if they are favourably oriented he radar beam, and in many strong permanent echoes, and it consists of pronounced maxima l minima. For the 10 cm range-height (RHI) radar at East Hill the first two lobes in the vertical ne are $2\frac{1}{2}°$ and $5\frac{1}{2}°$ from the centre of the main-lobe response, with minima at $1\frac{1}{2}°$ and 4°. follows that for a reflecting surface at 0·4° elevation there should be a main-lobe response bout 0·4° elevation angle, a first side-lobe response close to 3°, a second close to 6°, etc. With ny echoes on the display there should then be a series of bands radiating from the centre of RHI display, with pronounced minima between them. These angles may seem small, but it st be remembered that in the conventional RHI the scale of height is much more open than scale in range, with the result that a first side-lobe band would be *displayed* at about 20°, l a second at about 30°, which should make them more readily obvious on the display. None of se effects has been observed. Indeed, as he reports, the echoes are ' confined usually to a izontal belt,' irrespective of elevation angle.

(b) Further, if the motion of the reflecting surface is horizontal, then any apparent movement range should be along these sloping bands. In fact, the movement of high-elevation ' angels ' lmost always horizontal, which would be quite unreasonable on side-lobe theory.

(c) It should also follow that an obstruction subtending an angle of elevation greater than °, such, as a range of hills, would cut out all ' angels ' beyond twice the range of the hills. As a case in point, the Dunstable Downs, 2 to 3 mi distant, obstruct the East Hill radars to vations of 1°, yet no such reduction is observed. The high-elevation echoes appear no different his sector from those in any other sector.

(d) Finally, would not these reflecting surfaces, moving at heights of a few tens of metres ve the ground at appreciable speeds, be destroyed immediately by turbulence ?

These objections, in my opinion, make side-lobe explanation implausible, and since the total-ection mechanism cannot explain echoes above 0·4° elevation (which accounts for a very ge proportion of the total), I conclude that atmospheric inhomogeneities are a most unlikely se of radar ' angels.'

I suggest that there is now reasonably firm evidence that radar echoes of this type are caused the movements of birds (Harper 1958), and it is significant that no widespread displays have n seen outside the migration seasons. Indeed there is increasing recognition that studies of at interest in bird migration can be derived from them (e.g., Sutter 1957, Lack 1958).

There may be other unexplained echoes for which a meteorological explanation may yet ve valid, e.g., the echoes at sea-breeze fronts reported by Atlas *et al.* (1953), but Mr. Jones ws us clearly that great difficulties can be expected to arise in propounding them.

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5 December 1958.

Reply

By R. F. Jones

Although there is increasing recognition that studies of great interest in bird migration can be derived from the radar observations called ' angels,' I feel that a note of caution should be struck. Mr. Harper and the ornithologists may be right, but the possibility of an alternative explanation should not be overlooked, and Mr. Harper's objections in his letter are by no means sufficient to rule it out. With regard to (a), Table 2 of my paper shows that at short ranges reduction of transmitted power by as much as 60 db may be capable of giving an echo. I am sure that the radiation pattern of the East Hill height-range radar is not known sufficiently to be able to say that the minima in the side-lobe pattern are more than 60 db down and the method of display combined with beam-width spread does not show up intensity variations well. Even with a vertically-looking radar it is necessary to guard against deception by echoes from ground-based targets received from radiation transmitted at 90° to the main lobe. Objection (b) is presumably based on other observations since movement on a height-range radar is difficult to observe. Isolated targets followed by GL3 radar at short range are not necessarily typical of the mass of ' angels ' sometimes seen on height-range radar. It is not true, as stated in (c), that an obstruction of greater than 0·4° in elevation would cut out all ' angels ' beyond twice the range of the obstruction. ' Angels ' would still be visible, as shown by Eq. 11, by reflection of the main lobe at angles greater than the critical angle. Furthermore, since I only wished to establish a possible mechanism, I deliberately avoided consideration of the effects of a slight inclination to the horizontal of the reflecting surface. It is clear from simple geometry that you can get a wonderful variety of echo heights, ranges and speeds with very small angles of inclination ! Thus a reflecting surface 30 m high at a range of 2·7 km tilted towards the transmitter at an angle of 0·3° to the horizontal would give an apparent echo at a range of 21 km and height of 230 m and its apparent speed of movement would be 50 times that of the reflecting surface. There can, of course, be no firm answer to (d) but the meteorological conditions for a reflecting surface are those associated with extreme stability in the layer, and the ' angel ' echoes may indeed be evidence that such layers are not immediately destroyed by turbulence.

I think there can be little doubt that, if the requisite gradients of temperature and vapour pressure occurred, echoes similar to ' angel ' echoes would be seen. If the echoes were from isolated targets such as birds it is difficult to see why they were not equally detectable by the 3 cm wavelength height-range radar at East Hill which has comparable performance on such targets. It would seem wise, therefore, to keep an open mind on their explanation until direct observations on a sufficient scale resolve the problem.

Meteorological Office,
Air Ministry,
Kingsway, London.
1 March 1959

551.521.32 : 551.521.61

Absorption and emission in the atmospheric window from 770 to 1,250 cm^{-1}

By W. T. ROACH and R. M. GOODY

(Read 18 March 1959. Presented by Dr. W. T. Roach. See *Q.J.*, **84**, p. 319)

Dr. J. T. Houghton : Is it possible that further studies of the window in the atmosphere may be sufficiently accurate to distinguish between line-shape profiles given by different theories ?

Mr. F. Saiedy : I have made measurements of background extinction coefficient β in the window at 11·1 μ, using a spectrometer capable of resolving the individual spectral lines (resolution of 0·4 cm^{-1}). I have correlated β with the water vapour in the zenith deduced from scanning a single line of water vapour. I found 91 per cent correlation coefficient between these two variables. Simultaneously I measured the extinction coefficient in the visible (β_{vis}). Preliminary results show no significant correlation between $\beta_{11\cdot1\,\mu}$ and β_{vis}, which is in keeping with Elsasser's suggestion that background extinction is due to the accumulated wings of strong lines. Stated otherwise, these results show that $\beta_{11\cdot1\,\mu}$ may be as little as 0·03 when β_{vis} is as large as 0·3. These results are clearly inconsistent with those reported by Roach and Goody in their Table 4, range 5 – where an increase in $\beta_{11\cdot1\,\mu}$ of approximately 0·3 is recorded between Ascot and London – unless there is a different mechanism of extinction to be assigned to London haze.

Dr. A. W. Brewer : I would like to ask why the continuum seems to be so low in the atmosphere if it is due to the water vapour which is rather more uniformly distributed.

Dr. T. W. Wormell : The computed extinction curves for 3 μ water drops, and for water vapour absorption, are such similar functions of frequency that it is difficult to decide which is the better fit with the observations. Is there any simple physical reason for this remarkable similarity or is it just bad luck ?

Dr. Roach only discusses water drops of 3 μ radius. Would a range of sizes give the same effects ?

Dr. W. T. Roach (*in reply*) : Replying to Dr. Houghton I would say that improved observations of the continuum might reasonably be expected to clarify the situation, but it must be remembered that while disagreement between theory and observation enables us to reject the theory, the converse is not true : the results presented in this paper agree well with the Lorentz theory, but all we can say is that the resultant of the factors which modify the Lorentz shape far from the line centre must be small.

In reply to Dr. Brewer, the scale height of water vapour density is under 2 km. The effect of pressure on absorption lowers this value. Effective emission levels for water vapour are therefore between 1 and 1·5 km above ground level. This agrees with the observed continuum emission temperatures, as Fig. 6 of the paper demonstrates quantitatively.

Replying to Mr. Saiedy, the results as stated by Mr. Saiedy are not inconsistent with our hypothesis *as published* since we account for the increase in $\beta_{11\,\mu}$ in London by the not unreasonable suggestion that it is due to the greatly increased pollution there which has, however, a negligible effect at Ascot. Since β_{vis} was not measured in London, comparison of β_{vis} with $\beta_{11\,\mu}$ at Ascot does not affect this argument.

However, re-examination of the London results show that the observed proportionality between the logarithm of the deflection and the square root of the air mass (see p. 324 of the paper) could also have been produced by a small increase in the matter responsible for producing continuum extinction during the period of observation, *thus leading to an overestimate of* β. This is reasonable since the London runs were made after midday when convection and evaporation would be active, whereas the Ascot runs were made in the early morning and ceased by the time convection became vigorous.

The effect of this on our conclusions is to weaken the suggestion that aerosol extinction is large in industrial and city areas although it may still be detectable. No definite conclusion can be reached on this from the results at present available, and we await the results of Mr. Saiedy's observations, which are more precise than our own.

Reply to Dr. Wormell (*partly communicated*) : There is a broad similarity between the shapes of the water vapour and liquid water absorption spectra although the latter is about 10^4 stronger than the former. This is because while the rotational and vibrational transitions of the water molecule still occur, the resultant absorption spectrum is profoundly modified by the greatly increased collision rate and force interaction of the molecules due to their much closer proximity in the liquid state.

When we made our computations in terms of what is known about the optical properties of water droplets and the size distribution of aerosol in the atmosphere, we found nearly all the extinction would be produced by droplets in the range 1-10 μ radius since droplets smaller than 1 μ are small compared to the wavelength considered while the concentrations of droplets much larger than 10 μ become vanishingly small. Hence we found that the use of a monodisperse model of 3 μ droplets was adequate. As dynamical meteorologists are aware, one can often use the crudest of models to gain physical insight into a particular problem.

The increase of extinction with wavelength predicted by the model is due to the sharp increase in the absorption index of water below 900 cm^{-1} which swamps the normal decrease in extinction from a given particle distribution as the wavelength increases.

551.558.1

The motion in and around isolated thermals

By BETSY WOODWARD

(Read 13 May 1959. Presented by Mr. P. M. Saunders. See *Q.J.*, **85**, p. 144)

Before Mr. Saunders presented the paper, the development of ideas leading up to it was described by Dr. R. S. Scorer. The substance of his remarks is contained in two papers : ' Experiments with convection bubbles ' by R. S. Scorer and C. Ronne, *Weather*, May 1956, p. 151, and ' Experiments on convection of isolated masses of buoyant fluid ' by R. S. Scorer, *J. Fluid Mech.*, **2**, pt. 6, p. 583, Aug. 1957.

Mr. R. F. Jones : Dr. Scorer has described some fascinating experiments on the sinking of a heavier fluid through a lighter one with which it can mix; he describes the resulting movements as ' thermals.' How far does he believe these experiments are related to convection in the atmosphere ? In the experiments the density difference between thermal and ambient fluid is decreasing regularly with height; in the atmosphere the density difference may be very variable; in particular there is the contribution of latent heat to buoyancy in cloud. The experiments may be reasonably related to dry thermals but the extension of the theory by Mr. Saunders to movement in cloud seems to be a doubtful step to take.

Dr. R. J. Murgatroyd : I share the doubts of Mr. R. F. Jones concerning the general applicability of Dr. Scorer's model experiments to natural conditions. Our observations at the Meteorological Research Flight indicated that from near the surface to 1,000 ft or so, conditions were very disorganized. Above that level organized thermals of small dimensions (*c.* 100 m across) were found. Just below the cloud base the lapse rate became rather stable and there were indications that the convection was changing its scale to produce cells of the order of 1,000 m across, as often found in cumulus clouds. Since neither the effects of latent heat nor the continued production of suspended matter are catered for in the laboratory experiments, presumably they should be regarded mainly as applying to thermals well above the surface yet well below the cloud base (although the photographs shown make it tempting to identify the clouds in the laboratory experiment with cumulus clouds). Several other features of the model appear to be in good agreement with observation, e.g., the stabilization of the lapse rate just below clouds would be expected from the circulation scheme described.

Another point of considerable interest is that of the water content to be expected in a cloud with the circulation shown. Many measurements, usually of poor quality, have been made in the past which suggest that water contents as large as the adiabatic value are seldom found in cumulus clouds unless there is precipitation. Both lateral mixing (entrainment) and vertical mixing have been proposed as explanations for this. In this model neither of these effects appear to be important and it therefore seems that in the centre of a young medium-sized cumulus

should expect the water content to approximate to the adiabatic value. In my opinion this is bably correct and adiabatic water content would be found more often if adequate techniques measurement were available. I would be interested to hear Dr. Scorer's comments on this l also to see a computation of the water-content distribution in an average temperate-latitude mulus cloud based on this model.

Dr. R. C. Rainey : While the fact that these laboratory experiments have provided a model ich is consistent in general terms with gliding experience is certainly encouraging, the only dence from gliding which is actually quoted, either in this paper or in the references cited, pears to be distinctly subjective. Detailed analyses of the records of particular gliding flights ve in the past provided evidence, of a much more objective nature, on the strength and tribution of the vertical components of air motion : has objective evidence of this type, on the tailed relevance of this model to thermals in the atmosphere, yet been secured – or sought ? pre specifically, I should like to enquire what criteria were used by the author in deciding, m flight observations, that this model form had in fact been attained, at, for example, a given itude.

Dr. J. S. Turner : I have made some experiments similar to those of Dr. Scorer. The ference is that I have projected buoyant fluid upwards with high velocity, and have thereby oduced buoyant vortex rings having a larger and controllable circulation (in the mathematical nse); whereas the only circulation in the special case of the thermal has been produced entirely the action of the buoyancy in the early stages of the motion. As a buoyant vortex ring rises rough uniform surroundings it grows, more slowly than a thermal; but the circulation remains nstant, because one can always take a circuit round the buoyant, vorticity-containing core nich lies wholly in irrotational fluid of the same density. A thermal is slightly different, in at it has distributed, rather than concentrated vorticity and buoyancy; but does not Dr. Scorer ree that Miss Woodward's diagrams show that external fluid is being drawn up the centre of a ermal from behind, so that the circulation will remain constant in this case too, and for the me reason as it does in vortex rings ?

Mr. H. H. Lamb : It should be interesting to make use of the pattern and magnitudes of rtical motion in and outside cumulus and cumulonimbus cloud to estimate just how misleading individual radiosonde ascent can be. The possibilities of misleading ascents are probably their worst where the convection cloud pattern is stationary or moves only slowly over the land-ape. I remember a case in Malta where there was quite large cumulus (10,000 to 12,000 ft gh) and drizzly showers all afternoon over the centre of the island, associated with the convergence the sea breezes from opposite coasts; yet the radiosonde ascent made at Qrendi over the uthern shore, only 5 miles from where the showers were occurring, indicated pronounced bsidence to about 3,000 ft. Such difficulties may be serious in parts of the world where great stances separate neighbouring radiosonde stations. 1,000-500 mb thickness may be affected, well as the stability indications of the ascent. Can Dr. Scorer give any idea of the possible agnitudes of these effects ?

Sir Graham Sutton, F.R.S. : One of the most interesting results was that which related e diameter of the cloud to its vertical travel. The factor of proportionality was the same as at which enters into von Kármàn's well-known relation for the turbulent mixing-length and e height above the solid surface. Some years ago, when studying the dynamics of a small nvective plume rising in calm air in the laboratory I pointed out that the shape of the plume uld be calculated with considerable accuracy if one followed the main lines of the mathematical eatment of large-scale smoke clouds diffusing in a horizontal wind in a neutral atmosphere d using the same values of the diffusion coefficient. In other words, although there was an normous difference in scale between the tiny laboratory plume and the smoke cloud, the ocesses of growth appeared to be governed by the same laws, even to the magnitude of the ffusion coefficient. Now cloud physics has produced the same result. This, to say the least, very remarkable.

Dr. R. S. Scorer (*in reply*) : In reply to Mr. Jones, the answer is simply that we have vestigated what happens when buoyant miscible fluid is released in otherwise undisturbed rroundings. The motion is always of the form assumed by our thermals and this can be derstood theoretically. We have shown that erosion occurs if the surroundings have sufficient able stratification, but we have no quantitative measurements of erosion. The release of heat condensation does not materially affect the problem except that it is, of course, extremely fficult to imitate experimentally. However, the rise of thermals within a large cloud in which e lapse rate is wet adiabatic will be the same as that of a thermal in neutral surroundings except r the possible interference of other motions.

It is important to say that we have not attempted to explain everything, and observations which do not appear to fit in, or additional complications, do not in any way detract from what we have found out about how a buoyant mass will rise on its own.

I cannot agree with Dr. Murgatroyd's statement that the (stable) sub-cloud layer can be predicted from what we have found out about thermals. It is to be explained only in terms of larger circulations and the stability of the air between cumulus clouds. For instance, if there were sufficient horizontal convergence I don't think there would be a stable sub-cloud layer at all, and this has nothing to do with the character of individual thermals.

The water content to be expected in a cloud thermal depends on how many times it has been turned inside-out in rising from the condensation level. If it has risen only about one diameter the air in the centre will have risen almost undiluted from the cloud base. If it has entrained exterior unsaturated air up the centre, the liquid water content will be very low. If it has risen in clear air but has been eroded, the air rising up the centre will be air from near the original ring-shaped core and the liquid water content may be very nearly the adiabatic value.

The water content to be predicted in a particular cloud depends on all these considerations, and I think it accords fairly well with experience to say that anything, from almost nothing to the adiabatic value, can be provided by a capricious nature.

Dr. Rainey raises a most interesting question about how research on thermals should proceed. All along we have been in continual discussion with glider pilots experienced in the art of soaring in thermals and we find that the picture presented as a result of these experiments gives the most satisfactory interpretation so far obtained of their observations. But it is quite another matter to obtain and make use of aerial observations to further the accuracy of our picture. Miss Woodward has made several flights in a glider with sensitive (thermistor) pointer-reading thermometers but it has not proved possible to make any quantitative test of the theory. All it was possible to do was to give orders of magnitude to the temperature and humidity anomalies to be found in thermals.

The flights were valuable in that the observations could be fitted into the theory without difficulty; but I think laboratory experiments to measure density distribution, for example, in thermals would be more valuable. What gliders can do is to throw more light on the nature of thermals near to the ground, i.e. between their origin at the hot surface and their appearance as isolated thermals at a height of, say, 700 m. I would therefore advocate more laboratory rather than atmospheric experiments at this stage in our knowledge. Any analysis of the distribution of thermals based on traverses by powered aircraft such as Dr. Murgatroyd has in mind, is likely to be thoroughly misleading because the sampling of thermals is quite uncontrolled (we are not interested in statistics in this approach).

The most useful glider observations are those in which several gliders have been soaring in the same thermal, and their behaviour has fitted the laboratory picture very well. For instance those in the bottom of the thermal are usually able to soar up to those at the top but those a little below the bottom can make no progress at all. It is interesting that pilots used to think that the decrease in ascent rate experienced when a glider rises to near the top of a thermal, combined with an increase in the variability of the up-draught and the great difficulty in remaining in the centre (due to the outflow) was due to their falling out of the bottom of a thermal. One expression used by pilots, that they had " got to the top " of a given thermal (meaning that they had got to the point where they, personally, could make no further use of it) has turned out to be true in a sense they had not expected !

The advantage to be derived from making the most of soaring pilots' experiences is twofold. First it will prevent the theoretician from claiming too much – he will be saved from being naive by being continually reminded of the enormous complexities in nature. Secondly it will pose new problems to him. For example the problem of cloud streets and thermal streets, classically explained in terms of a viscous fluid, must be solved all over again now that we know that thermal motions are independent of viscosity.

I agree with Dr. Turner that the circulation remains constant in a thermal, but I believe this to be the case not because it is observed that exterior fluid passes up the centre but because it can be proved theoretically that it remains constant. The fluid does indeed pass up the centre but as it nears the top of the thermal it begins to wobble around. We made one experiment in which marked fluid was entrained into the mixing region at the front of the thermal but into only a small region well to one side of the axis. This fluid circulated round the ring-shaped core and was well mixed as it came up towards the cap so that it then passed round the thermal in all directions. If fluid entrained at the front on one side can make its second circuit round the other side then fluid entrained up the centre must be well mixed before it diverges outwards at the top.

If the exterior fluid did pass right up the centre then the circulation would remain constant, but it is possible that it gets mixed before reaching the cap and that the circulation generated by the smaller density at the cap is exactly cancelled by Reynolds stresses in the mixing region. I do not know the answer to that and it requires fuller investigation.

The possibilities of misleading radiosonde reports from inside cumulus clouds are at least partly indicated by the patterns executed by thermals and Mr. Lamb is invited to invent possible situations for discussion. Perhaps more notice should be taken of the rate of ascent of the balloons. But one interesting feature of thermals is that they are warmer (potentially) than the air above them : therefore if a balloon were to rise out of the top of a thermal it might record a super-adiabatic (' wet ' within a cloud) lapse rate, which would, of course, not be a general feature of the neighbourhood. If Mr. Lamb wants to estimate the magnitude of errors in thickness measurements he must first estimate the possible temperature excess in the thermal which carries the balloon up, and this depends on measurements in the atmosphere, not on the theoretical behaviour of thermals : therefore he is probably better informed than I am.

Though I have learned a great deal from Sir Graham Sutton, particularly in connection with thinking about convection, and I gratefully acknowledge my debt to him, I think it would be appropriate to point out here one of the differences in this approach to convection via individual thermals. This is that we do not consider or discuss anything in the nature of a diffusion coefficient. Such a concept, I think, is only appropriate when one is thinking of large numbers of thermals, that is, when one makes a statistical approach. The nature of the motion is so different in the surroundings and in different parts of the thermal that statistical concepts are irrelevant. We doubt the meaning of such an idea as a turbulence spectrum in a cumulonimbus : it only applies to hundreds of clouds taken together. The statistical approach is, of course, legitimate and profitable in certain contexts, but it is not the approach we have made, and I do not feel competent to comment on it.

551.509.313 : 551.511.3 : 551.551 : 551.558.21

Numerical experiments in the integration of the meteorological equations of motion

By E. KNIGHTING, D. E. JONES and M. K. HINDS

(Read 17 June 1959. Presented by Mr. E. Knighting. See *Q.J.*, **84**, p. 91)

The introduction of the effects of topography into methods of numerical forecasting

By J. S. SAWYER

(Read 17 June 1959. See *Q.J.*, **85**, p. 31)

The above two papers were presented at the Ordinary Meeting in the Society's rooms. Following this Mr. G. A. Corby gave a brief account of the first experiment in numerical forecasting carried out on the electronic computer recently installed in the Napier Shaw Laboratory, Dunstable. The experiment comprised a series of forecasts using the Sawyer-Bushby two-level model, commencing both from initial fields interpolated manually from conventionally analysed charts as well from fields computed by the machine from data extracted and analysed automatically. Some verification statistics were presented showing that the numerical forecasts were somewhat better at 500 mb than at 1,000 mb although at both levels the standard achieved differed little from that of the corresponding conventional forecasts. The meeting was then opened for general discussion.

DR. R. S. SCORER : I would like to express a viewpoint about the use of numerical meth and machines in the dynamical problems of the atmosphere. There is no possibility of prov a general result by treating a few, or even many, actual weather situations. There seems to evidence of too little general theoretical thinking and too much messing around comparing res with actual weather developments and with other kinds of forecast. If we find that different meth of grid-making or slightly different boundary assumptions or changes in time interval, etc., h effects difficult to understand they should be studied by applying them to special simple artiff situations – the actual developments of real situations have nothing whatever to do with problem. It may be quite a big problem to find the special artificial situations in the first pl but a major effort should be put into finding them because the problems will not be propo solved otherwise.

PROF. H. G. FLEAGLE : In considering the significance of Mr. Knighting's results i important to know whether the differing results which he has described are to be attributec change in the grid size used or to change in the orientation of the grid points. One must exp use of a larger grid size to yield ' forecasts ' with less detail than those with smaller grid s On the other hand, a systematic asymmetry in pressure fields which would make ' forecas generally sensitive to orientation of grid points would be surprising and somewhat distressi I should like to ask Mr. Knighting whether tests were conducted using two orientations but same grid size.

(*Partly communicated*) : I should like to support Dr. Scorer's point of view, although appears to be in no need of assistance in this discussion. In order for the work in numer prediction to contribute towards understanding of the behaviour of the atmosphere, as well towards a really useful and reliable forecasting method, we must understand the properties the models being used. This can be achieved more efficiently, it seems to me, by conduct many experiments using the same idealized initial state than by accumulating a series ' forecasts ' from current data, changing the model as defects are recognized, and finally judg the efficacy of the final model by comparing correlation coefficients or by majority vote.

We can learn from the experience of earlier experiments in numerical prediction, and particular from the experience of the Joint Numerical Weather Unit in Washington. H several years of daily numerical predictions using current data have shown that the barotro model does about as well as the judgment of the human forecaster – this we knew pretty v in 1951 or earlier. The baroclinic models tested appear to be no better, if they are as good, the barotropic model, and this is disappointing. But it seems to me much more disappoint that, despite the efforts of a highly capable group, we have learned so little about the physic large-scale atmospheric behaviour beyond what we knew before the Joint Numerical Weat Prediction Unit was formed. I think carefully designed numerical experiments can tell us a gr deal about the atmosphere that we can learn in no other way, and I hope that the Dunsta group will be able to devote its best efforts to this type of investigation.

MR. A. G. MATTHEWMAN : Some major ' mechanical ' errors are of a comprehensible ty such as orographic errors, and could be more or less corrected on the lines of Mr. Sawyer's pa but more fundamental errors naturally occur because of failures of the basic approximatio Sometimes a thickness pattern is changing only slowly, or is accompanying a system with li change, and although it appears favourable for, say, deepening a depression, the ' convention forecaster knows that recently the depression has not been deepening, has perhaps even fill accordingly, he probably weights this history and forecasts the depression not to deepen a perhaps to fill. The machine, however, deepens the depression and thereby probably produ an inferior forecast.

It would seem advantageous (at least until the mechanical forecasts are clearly better) to h a backcast as well as a forecast produced by the machine from each set of data, produced by taki opposite signs of t, time. At least in some situations, comparison of the backcast with the act earlier chart might show up the sort and size of errors likely to be occurring in the mechani forecast.

MR. L. C. W. BONACINA : I should like to ask Mr. Sawyer whether he is hopeful that d day it will be possible to predict orographic components of rainfall provided the general amou of precipitation due to an oncoming depression or front could be foreseen and provided, al the reactions of a particular region or locality to prevailing and other wind directions w known.

Upon this possibility depends the further possibility of producing a map of the land surf of the globe marked off with plus or minus (rain-shadow) orographic factors, including second local-plus effects in a minus region and local-minus in a plus region.

It always seems to me that a travelling rain system is not a complete entity unless related to the surface over which it passes – all up in the air so to speak. I am not quite clear whether it would be theoretically sound to regard the ocean surface as a basis of comparison for the land surface, for although there are no orographic effects over the open sea the comparison might be vitiated by the very different conditions of evaporation and convection.

Mr. E. Knighting (*in reply*) : I agree with Dr. Scorer that numerical experiments should be carried out to investigate truncation errors and in fact the paper under discussion was designed to reveal the magnitude of the problem. Theoretical work has been carried out on this problem in all main Numerical Weather Prediction centres. Such experiments cannot be carried out on an artificial field at the moment because we have not yet discovered an analytic solution of the equations which retains the characteristic synoptic features. In practice the truncation error cannot be divorced from the errors of interpolation in the initial field nor from the wavelength of the features which the equations are designed to deal with. Some experiments with analytic initial fields are being carried out by Mr. Sawyer. Professor Fleagle has put his finger on a problem that ought to be dealt with when he asks if experiments have been made with the same grid size but with different orientations ; we have not made such experiments. I share his disappointment but not his pessimism in the progress of numerical weather prediction. He is, however, unfair to baroclinic models which give information throughout the lower troposphere whereas the barotropic models give information only at the mid-level. Barotropic forecasts at, say, 500 mb are as good as, or better than, baroclinic forecasts at this level because the former proceed from a stream-function field and so far no exhaustive tests have been made of a baroclinic model incorporating a stream-function technique. It is not our physics which is wrong; the mathematics have not yet been worked out.

Mr. J. S. Sawyer (*in reply*) : As Mr. Matthewman points out, errors in numerical forecasts are likely to be systematic and to some extent predictable from the behaviour of the equations applied to a ' backcast ' or from statistics of previous forecasts. Here is a possibility of improving numerical forecasts without gaining more physical insight. This procedure would be the opposite of that advocated by Dr. Scorer and so far we have not explored it at Dunstable.

In reply to Mr. Bonacina, the vertical motions which produce rain, particularly orographic rain, have a smaller horizontal scale than the systems ordinarily treated in numerical forecasting. Detailed prediction of the rainfall pattern would therefore require a different formulation of the mathematical problem than that at present used and tested. It also requires knowledge of the distribution of water vapour and cloud in some detail. A simple extension of current methods of numerical forecasting will therefore not be suitable for predicting details of the rainfall distribution although Dr. Smagorinski in U.S.A. has had some success in predicting the average rainfall over areas some hundred or more miles in extent.

I do not think the problem of calculating orographic rainfall is a completely impossible one, but much theoretical and mathematical work remains to be done. As the effort needed is great, I am not sure that it will attract the necessary interest. It will certainly be many years, probably decades, before Mr. Bonacina sees results such as he has in mind.

Proceedings at meetings of the Society

Special General Meeting, 17 June 1959

A Special General Meeting of Fellows was held to transact the following business :

Item 1. To elect the Officers and Council for the year 1959-60.

Fellows had been circulated on 27 April with the list of nominations by Council for Officers and Council for the year 1959-60. No other nominations having since been received, the persons so nominated were declared to be elected unopposed in accordance with By-law 51.

The Officers and Council for 1959-60 were therefore elected as follows :

PRESIDENT

JAMES MARTIN STAGG, C.B., O.B.E., M.A., D.Sc.

VICE-PRESIDENTS

RICHARD MEAD GOODY, M.A., PH.D.
HOWARD LATIMER PENMAN, M.Sc., Ph.D., F.Inst..P.
GEORGE DAVID ROBINSON, Ph.D., F.Inst.P.
PERCIVAL ALBERT SHEPPARD, Hon.A.R.C.S., B.Sc., F.Inst.P.

VICE-PRESIDENT FOR CANADA

WARREN LEHMAN GODSON, M.A.

VICE-PRESIDENT FOR SCOTLAND

ROBERT CRANNA, M.A., B.Sc.

TREASURER

BRIAN CECIL VERNON ODDIE, B.Sc.

SECRETARIES

ARTHUR GEORGE FORSDYKE, Ph.D., A.R.C.S. (Librarian)
PATRICK JOHN MEADE, O.B.E., B.Sc.
JOHN STANLEY SAWYER, M.A. (Editor)
RICHARD SEGAR SCORER, M.A., Ph.D.

COUNCILLORS

JOHN ALAN CHALMERS, M.A., Ph.D., F.Inst.P.
GEORGE ARTHUR CORBY, B.Sc.
JOHN ROBERT DARK FRANCIS, B.Sc. (Eng.), A.I.C.E.
FRANK HENRY WINN GREEN, M.Sc.
ARTHUR LEONARD MAIDENS, B.Sc.
ROBERT JAMES MURGATROYD, Ph.D.
PAUL ROBERT OWEN, B.Sc., A.F.R.Ae.S.
REGINALD CHARLES RAINEY, Ph.D.
JOHN BRODRICK RIGG, M.A.
JOHN RUSSELL THORP, O.B.E., M.A.
CHARLES DESMOND WALSHAW, M.A., Ph.D.
HENRY LIONEL WRIGHT, M.A.

Item 2. To amend the Society's By-laws in accordance with the Notice of 11 May sent to all Fellows :

(a) To abolish the grade of Associate.

(b) To offer a concessionary rate of subscription to Fellows residing outside the British Isles.

To give effect to these changes the following amendments to the By-laws were proposed :

1. By-law 5. Delete ' Associates '
2. By-law 7. Delete ' Associates '
3. By-law 10. Delete ' Associate ' in line 1 and ' Associates ' in line 6
4. By-law 11. Delete ' Associates '
5. By-law 16. Delete
6. By-law 21. Delete ' or Associateship '
7. By-law 35. Delete ' Associates £1. 1s.'
8. By-law 36. Delete ' Associates £3. 0s.'
9. Insert new By-law 37. ' Fellows residing outside the British Isles may if they so desire pay a reduced annual subscription of 75 per cent of the full rate.'

The resolution was carried unanimously.

Ordinary Meeting, 17 June 1959

The following candidates were declared to have been elected on 22 April by the Council :

FELLOWS

ADKINS, Clement John, M.A., Ph.D., Jesus College, Cambridge (*transfer from Student Membership*).
BANYARD, Capt. Richard Stansbie, A.M.I.E.T., White House, Chapel Road, Tiptree, Essex.
BROWN, Instr. Lt. Neville George, B.Sc., B.A., R.N., Treetops, Hill Road, Watlington, Oxfordshire.
CHORLEY, Richard John, M.A., Department of Geography, Downing Place, Cambridge.
CLARK, Richard Peter Kelvin, B.A., Ford Castle, Berwick-on-Tweed, Scotland.
HEPWORTH-NICHOLSON, Simon, B.A., Trinity College, Cambridge (*transfer from Student Membership*).
MATTHEWMAN, Albert Gerald, 30 Church Crescent, St. Albans, Herts. (*reinstatement;* resigned February 1958).
MCDONALD, Flt. Lt. Angus James, Department of Flight, College of Aeronautics, Cranfield.
SCOTT, George, 4 Bastle Road, Foulden, Berwick-on-Tweed, Scotland.
STRINGER, S. Winifred, Roxburgh, Colwall, Malvern, Worcester.

STUDENT MEMBER

VINCER, Michael, 136 Tubbenden Lane, Orpington, Kent.

The following candidates have been proposed for election under By-law 25 :

Foreign Members

Cantor, Israel, A.B., M.S., 17 Arborway, Indian Springs Estates, Asbury Park, New Jersey, U.S.A.

Chiu, Wan-Cheng, B.S., M.S., Ph.D., Department of Meteorology and Oceanography, New York University, New York 53, N.Y., U.S.A.

Mourabet, Abdul Rahim, M.Sc., B.Sc., c/o S. D. Mourabet, Fardous Street, Tarazi Building, Damascus, U.A.R. (*transfer from Student Membership*).

Canadian Branch

Associate Members

Belhouse, Henry C., M.A., c/o Meteorological Office, Vancouver Airport, British Columbia.

Bernachi, D. A., B.A., c/o Officers' Mess, R.C.A.F. Station, Trenton, Ontario.

Bolingbroke, Dennis Oliver, B.A., A.F.O., R.C.A.F. Station, Cold Lake, Alberta.

Coté, Jacques, B.A., Officers' Mess, R.C.A.F. Station, Trenton, Ontario.

Dumka, Ernest A. William, B.A., Officers' Mess, R.C.A.F. Station, Trenton, Ontario.

Hone, Robert Stewart, M.A., 4865 Maplewood, Apt. 8, Montreal, Quebec.

Janecek, Paul, B.A., c/o Officers' Mess, R.C.A.F. Station, Trenton, Ontario.

O'Brien, Ralph Harvey, B.A., 1 F Wing, R.C.A.F., Capo, 5052, Canadian Armed Forces, Europe.

Pearce, John Charles, M.A., 1516 Pine Avenue, West, Montreal 25, Quebec.

Raniowski, Antony, B.A., 287 Ossington Avenue, Toronto 3, Ontario.

Rayner, John Norman, B.A., Geography Department, McGill University, 539 Pine Avenue W., Montreal, Quebec.

Shimizu, Gordon Masahiro, M.A., D.A.F.O., Domestic Terminal Building, Montreal Airport, Quebec.

Sparrow, David Murray, B.A., c/o Officers' Mess, R.C.A.F. Station, Trenton, Ontario.

Treidl, Rudolph A., M.A., Ph.D., 5 Brookmere Road, Rexdale, Ontario.

Tyner, Ralph Vernon, M.A., 3855 Sunnycrest Drive, North Vancouver, British Columbia.

Van Leeuwen, H. C. T., B.A., Aviation Forecast Office, R.C.A.F. Station, Moose Jaw, Saskatchewan (*transfer from Fellowship to Associate Membership*).

Wyllie, William David, M.A., c/o D. O. T., Goose Bay, Labrador, Newfoundland.

Fellow

McQuarrie, A. F., B.A., Aviation Forecast Office, Municipal Airport, Calgary, Alberta (*transfer from Associate Membership to Fellowship*).

The following were declared to have been elected on 20 May by Council :

Fellows

Andrews, John Browning, B.Sc., A.R.C.S., 68 Friars Street, Sudbury, Suffolk (*transfer from Student Membership*).

Hanson, Flt. Lt. Colin Morris, No. 75 Squadron, R.N.Z.A.F., Royal Air Force Tengah, Singapore 24.

Hickson, Peter, M.A., 12 Witaka Street Sth, Lower Hutt, Wellington, New Zealand.

Millar, Gordon Woods, 47A Heriot Row, Edinburgh 3, Scotland.

Nixon, Eric, B.Sc., F.S.S., 1 Detillens Lane, Limpsfield, Oxted, Surrey.

Rantzen, Henry, M.I.E.E., B.Sc., S.M.I.R.E., 12 Hocroft Road, London, N.W.2.

Roake, Harold Stuart, c/o James Howorth & Co., Victoria Works, Farnworth, Bolton, Lancs. (*transfer from Associate Membership*).

Student Members

Power, Christine Anne, 51 Larksfield Avenue, Throop, Bournemouth, Hampshire.

Shepherd, Margaret Olive, 56 Summit Drive, Woodford Green, Essex.

Smith, Brian Philip, 3 Holyrood Terrace, Holly Road, Westham, Weymouth.

Williams, Anne, 22 Orford Road, Walthamstow, London, E.17.

Willson, Robin Linhope, 17 Cloverfield Avenue, Fawden, Newcastle upon Tyne 3.

The following candidates were declared to have been elected that day by the Council :

Fellows

Buvanasundaram, Visuvalingam, 17 Queen's Gardens, London, W.2.
Donovan, B. J., F.R.I.C., 35 Gorsehill Road, Wallasey, Cheshire.
Harrison, Frank Robin, B.A., B.Sc., 431 Hillcross Avenue, Morden, Surrey (*transfer from Student Membership*).
MacDowall, Joseph, M.A., 21 Raisins Hill, Pinner, Middx.
Rewari, Sham Sunder Singh, I. Sc., Third Officer, M/S India Steamship Co. Ltd., 21 Old Court House Street, Calcutta 1.
Spolton, Lewis, B.Sc., M.E.D., 12 Bourne Avenue, Kirby-in-Ashfield, Nottingham (*reinstatement;* resigned 1958).
Stubbs, Raymond Frederick, 10 Beechlands, Bishop's Stortford, Herts.

Foreign Member

Inoue, Eiichi, National Institute of Agricultural Sciences, Section of Physics and Statistics, Nishigahara, Kita-Ku, Tokyo, Japan.

Canadian Branch

Fellows

Arnold, George Waddington, 179 Lawrence Avenue East, Toronto 12, Ontario.

Associate Member

Bourke, Roy Stuart, B.A., Central Analyses Office, Meteorological Branch, Trans-Atlantic Administration Bldg., Montreal Air Port, Quebec.

The following papers were read and discussed :

1. ' Numerical experiments in the integration of the meteorological equations of motion ' by E. Knighting, D. E. Jones and M. K. Hinds (*Q.J.*, **84**, p. 91).

2. ' The introduction of the effects of topography into methods of numerical forecasting ' by J. S. Sawyer (*Q.J.*, **85**, p. 31).

A discussion on numerical forecasting was then opened by Mr. G. A. Corby.

Summer Visit, 8 July 1959

The Summer Visit of the Society was held on 8 July at the Hydraulics Research Station of the Department of Scientific and Industrial Research at Howbery Park, Wallingford, Berkshire, by kind permission of the Director, Sir Charles Inglis.

Some 30 members attended, and, after an introductory address by Mr. N. P. Radley outlining the history, organization and work of the Station, the party was shown round the Main Hall which houses the laboratories and workshops, and saw several models where research was being carried out on behalf of various public departments on specific problems, such as prevention or reduction of flooding. In the grounds of the Station, model investigations of harbours in the wave basins were inspected.

At the conclusion of the visit Dr. R. S. Scorer moved a vote of thanks and expressed appreciation of the excellent arrangements made by Mr. Radley, and for the services of the guides.

A full account of the Visit appeared in the September issue of *Weather*.

Reviews

Physical climatology. By H. Landsberg. Pennsylvania (Gray Printing Co.), 2nd edition, 1958. Pp. xi, 446, 108 Figs.; 224 Tables. $6.

This work tries to bring the first edition (1941) up to date. Being an ' elementary textbook on physical climatology ' (preface, *iii*), its scope is vast, its level elementary and therefore much generalization of data and personal selection of material has been involved.

Chapter 1 deals effectively with the methods of climatological investigation, the installation of instruments and simple statistical treatments of climatic data. Chapter 2 considers, with varying success, the atmosphere and the elements of climate. Chapter 3 tries to embrace the whole climatic spectrum from global to ' spot ' climates, and in Chapter 4 an attempt is made to cover the enormous field of applied climatology.

The author should be taken to task immediately for certain terminological short-comings. Firstly, climatology is placed within the field of meteorology (preface, *iii*), then classified as a ' neighbouring field ' (introduction, *viii*) to meteorology, and finally (introduction, *viii* and *ix*) such attempts at rigid definition are discouraged because of ' the fluid state of this technical field.' This confusion is strikingly reminiscent of the contemporary disputation on this side of the Atlantic concerning the relative roles of geographers and meteorologists in studies of the atmosphere. Surely the proper reconciliation of this professional *impasse*, and the clarification of the text under review, would be derived from mutual respect for *methods* employed by each approach. Such respect, however, derives only from practical knowledge of methods, the mutual experience of which should be encouraged by geographers and meteorologists alike in related fields such as climatology. A second terminological inadequacy is the persistent misuse (e.g. *vii*, 158, 312, 332) of the geographical term ' topography,' which is *not* synonymous with relief or land surface but which refers to the general landscape of a place, including man-made as well as natural features. ' Influence factors ' (p. 233) is another most unfortunate expression.

A valuable collection of varied climatological data, some of it aptly illustrated, is available for reference, but more graphic and cartographic illustration would have been welcome. Again, some 400 references are listed enabling the reader to pursue a topic of special interest but, perhaps inevitably in an American work, references to relevant British papers and to certain cognate aspects of British Meteorological Office practice are relatively limited. For example, several ' growth ' thresholds are quoted. e.g., 41°F, 43°F (pp. 159, 382) but the value currently adopted by the British Meteorological Office (viz. 42°F) is ignored. The concept of the ' growth ' threshold is, quite justifiably, criticised (pp. 159, 382) by the author who, having done so, however, proceeds to overstress the precise ecological significance of the ' freezing ' threshold at 32°F (pp. 158-9). Furthermore, south-west facing slopes are *not always* the warmest (p. 332). The general problem of land aspect and insolation variables is most inadequately covered.

The absence of adequate reference to radiosonde equipment in Chapter 1, and the tendency for ' upper air ' to receive only incidental mention (e.g., p. *xi*) betrays the two-dimensional emphasis of this work.

The above criticisms apart, this book is a useful addition to introductory climatology texts and offers a wide range of reference data. In particular, it should educate British readers to the far-reaching implications of ' applied climatology ' in everyday life. It is a well-produced book but perhaps slightly over-priced.

J.A.T.

Advances in geophysics. Vol. 5. Eds. H. E. Landsberg and J. Van Mieghem. New York (Academic Press), 1958. Pp. ix, 325, Figs.; Tables. $10.

The fifth in this series of volumes, which review recent advances in geophysics, covers a variety of subjects, most of them being of topical interest.

It opens with an assessment, by N. C. Gerson, of the objects, organization and scientific programme of the Polar Years 1882-3 and 1932-3, and with a description of the aims and scope of their more ambitious successor, the International Geophysical Year. This article, although containing some interesting historical material, seems rather out of place in this volume, especially as it is so much more concerned with plans and organization than with the results of the scientific work.

The reviews on ' Microseisms ' by B. Gutenberg, on ' The size and shape of the Earth ' by R. A. Hirvonen and on ' The physics of cloud modification ' by J. E. MacDonald provide useful summaries for the non-specialist but, as far as the specialist is concerned, are lacking in penetration and critique.

The two most scholarly articles are those in which Dr. A. T. Doodson reviews observational and theoretical studies of oceanic tides and K. Watanabe discusses ultra-violet absorption processes in the upper atmosphere. This last article contains much valuable data and has an extensive bibliography.

B.J.M.

Once round the sun. By R. Fraser, London (Hodder and Stoughton), 1957. Pp. xv, 160; Figs.; Plates. 16s.

The title of this book should perhaps have been ' one-and-a-half times round the sun,' as it is a popular account of the International Geophysical Year which started 1 July 1957 and continued until the end of 1958. The author is the Administrative Secretary of the International Council of Scientific Unions and his aim has been to present to the widest possible audience the various aspects of the work achieved by international cooperation during this period. The first part of the book consists of five chapters explaining the main subjects

of investigation : the spinning earth, (gravity, rotation, magnetic field); climate and weather (condensed simple account of atmospheric constituents, radiation, global circulation and weather systems); currents, waves and tides; patterns in the outer atmosphere (photo-chemistry, airglow, ionosphere, tides, aurora); and sunspots and solar flares. The second part of the book describes the history, organization and work carried out during the I.G.Y. on the problems introduced in Part 1. Particularly informative are the chapters on Antarctica and rockets and satellites. At this time, however, the most interesting is probably the very short final chapter entitled ' summing-up – what has been gained from the work of the I.G.Y. ? ' Perhaps the greatest achievement has been that 54 nations have worked amicably together throughout this period on a common task. Practically, the author thinks it possible that its work may provide results leading to long-range weather forecasting, better radio communications and improved maps of the world, as well as imponderable advantages resulting from new fundamental geophysical discoveries. How far any of these hopes will be realised remains to be seen but it is very doubtful if the meteorological effort, for instance, could have been justified entirely by the first of these. Finally, the author concludes that the basic reason for it is that quoted from F. Nansen – ' Man wants to know, and when he ceases to do so he is no longer man.'

This book is of general interest to anyone desiring a quick, easily readable account of the I.G.Y., but of course it is not designed for more serious study. Although explanations are simplified and difficulties glossed over, no serious mistakes are noted. The production, with many figures and excellent, but now rather well-known plates, is very good and the price is quite reasonable.

R.J.M.

Physikalisch – Statistische Regeln als Grundlagen für Wetter-und Witterungsvorhersagen. Bd.2. By F. Baur. Frankfurt am Main (Akademische Verlagsgt.). 1st Ed. 1958. Pp. 152 : 23 Figs.; 61 Tables. 45DM.

In the present volume Dr. Baur seeks to establish rules having physical and statistical justification as a basis for forecasting the character of the weather over periods of a month or a season. The first volume gave about 60 such rules for short or medium-range forecasting, and the present volume gives about 40 more for the longer periods. As a fair example, I quote Rule 66 ' If November in Winnipeg is more than 3·5°F cold and in Boston the first half of December is cold or not more than 0·2°F warm, it will be highly probable that in Southern New England winter as a whole in temporal and spatial average will be colder than normal or in the extreme 0·9°F warmer than normal.' The evidence for this rule is given in Table 18, which names the 25 years in the period 1873 to 1954 in which the November temperature at Winnipeg was at least 3·5°F below normal and shows that in the 13 of these years in which the temperature at Boston in the first half of December was near or below normal the following winter in Southern New England was cold or near normal. Since the probability of a single winter chosen at random being cold or near normal is 0·6 then the probability of all of a sample of 13 chosen at random having this character is extremely small. Hence, the argument goes, the criteria used in choosing the sample, the preceding temperatures at Winnipeg and Boston, are of some assistance in finding the cold winters.

Following this pattern of investigation, Dr. Baur produces rules mainly concerned with large-scale patterns of the atmospheric circulation, proceeds to singularities in the annual pressure and temperature cycles, and finally discusses solar-terrestial relationships. In all cases the basic data are very fully set out, even when they do not entirely support the author's thesis, and a serious attempt has been made to distinguish fact from fancy by including only results which prima facie are unlikely to arise by chance. Nevertheless, the significance of the results may be overestimated, because the choice of variables between which the association is quoted seems to be dictated not by prior physical considerations, but by the magnitude of the association. In the example quoted, why is the New England winter temperature related to that at Winnipeg in October rather than September or November ? Presumably because the author either knew, from other work on the same data, or found through his own work, that the other months would give worse results. But if so, the association found must be tested for significance not as a lone result, but as the best of a considerable number of prima facie equally promising relationships which might have been published, and in these circumstances the threshold of chance occurrence is not only higher, but much more difficult to estimate.

While this consideration, which applies to most of the tables, reduces the value of the results, it should be said that nearly all work on long-range forecasting published to date is open to the same criticism, and that it is very difficult to see what Dr. Baur could have done about it.

This may be the last important book on long-range forecasting to be published by an individual scientist, rather than by a team of scientists working with an electronic computer, and for all its limitations it is an impressive record of the progress which one scientist has been able to make in studying an intractable subject.

J.M.C.

Recent advances in atmospheric electricity. Ed. L. G. Smith, London, New York, Paris, Los Angeles (Pergamon Press), 1958. Pp. xv, 631; Figures; Tables. £6.

This volume comprises the papers read at the Second Conference on Atmospheric Electricity, held at Portsmouth, New Hampshire in May 1958, sponsored by the Aerophysics Laboratory of the Air Force Cambridge Research Center, and includes some of the discussion. There are 48 papers and 10 ' short contributions.'

The participants in the Conference comprised most of the leading figures in atmospheric electricity at the present time, and, as each was asked to give a paper, it is not surprising that there is some unevenness in the contributions. Some workers had original results for which the Conference provided an opportunity of publication, while others were only able to present either a ' re-hash ' of previous papers, papers that were being published elsewhere, or interim reports of work in progress. Thus the prospective buyer of the book must not expect a logically planned survey of the whole field nor even exactly what would be conveyed by the title (better though this is than the title chosen for a similar collection of papers from another Conference a few years ago).

The presentation of the papers and diagrams and the general ' get-up ' of the volume is of the very high standard we expect from the publishers, but it seems open to question whether the bulk of the material is of sufficient permanent interest to warrant such trouble, which has, necessarily, put the price so high as to be prohibitive for any young worker in the subject. The method of publication of Conference proceedings as special numbers of current journals would appear to be a less costly method and one giving a wider circulation to those interested.

However, there are many of the papers which will be essential reading to those engaged in the subject and these will have to borrow the volume if they cannot buy it.

There is no index and the author index does not come up to the high standard of the rest of the volume; there are at least 12 mis-spellings of names and 2 authors have been separately listed with fictitious initials as well as with their own. In making the list of authors, the fact of a contribution to the volume is not considered a qualification, and not all those referred to are included; there is even, in one case at least, a reference to a blank page !

J.A.C.

Cosmic electrodynamics. By J. W. Dungey. London (Cambridge University Press), 1958. Pp. ix, 183; Figures; Tables. 32s. 6d.

Magnetohydrodynamics in its present stage of development requires textbooks which give lucid explanations of the derivation of the basic equations, adequate physical discussion of the effects of the extra magnetic terms, if I may be permitted to speak of the hydrodynamical equations as the basic ones which are modified by these additional external forces, and a summary of the results achieved up to the present. Dungey's book largely achieves these aims, but as one might have expected in such a small book the material is necessarily condensed.

The derivation of the basic equations using the Chapman and Cowling statistical treatment is done elegantly and simply, although not with the same detail as in the original. The qualitative discussion of the effects of the Lorenz force is clearly done although as a hydrodynamics man I would have welcomed mention of the analogy with vorticity in non-magnetic liquids. The special problems discussed range from the magneto-static, where the motion of the material is ignorable, to Alfven waves, the Chapman-Ferraro theory of magnetic storms together with a readable physical description of solar phenomena. It is possible that the book attempts to cover too wide a field since only short discussions can be given to such things as turbulence and shock waves, which from one's experience in hydrodynamics are almost certainly highly complex phenomena. The final chapter, devoted to ionospheric electrodynamics is the chapter which is likely to interest the meteorologist most, since it is unavoidably necessary to incorporate the ionosphere into our schemes of the general circulation of the atmosphere – unfortunately the basic equations to be used are even more formidable than those appearing earlier in the book. A summary is given in this chapter of Fejer's treatment of the tidal problem in the ionosphere and of the motion of irregularities in the electron density pattern.

T.V.D.

The Gulf Stream : a physical and dynamical description. By H. Stommel. London (Cambridge University Press), 1st Edition, 1958. Pp. xiii, 202 : 82 Figs.; 8 Tables; 3 Appendices. 45s.

The appearance of a book devoted almost entirely to a description of the Gulf Stream and covering the numerous advances made in our knowledge of this subject in the past two decades is a notable event for meteorologists as well as for oceanographers. The author is an oceanographer of international repute who has devoted many years of research to problems concerning the Gulf Stream. Meteorologists may regret that the scope of the book has been limited to the Gulf Stream system, as though it were in any sense separable from the general circulation of the North Atlantic Ocean. However Professor Stommel's argument for this treatment, even though justifiable only to an oceanographer, is that from an historic point of view the Stream has always been a primary feature of interest and scientific study in the Western North Atlantic because it was so clear-cut an entity, whereas over vast oceanic regions adjacent to it the current features are very faint and indistinct. Naturally many features and problems relating to the Gulf Stream which are described in this book, primarily concern oceanographers. However, certain physical characteristics of the upper layers of the oceans, notably surface current velocities and temperatures, and the depth of the thermocline, undoubtedly contribute to modify the overlying atmospheric flow patterns to some extent; and hence, portions of the book should have considerable value for meteorologists.

After starting with an excellent survey of the early explorations and ideas on the origin of the Gulf Stream by fishermen, navigators and scientific investigators, from 1700 onwards, the author proceeds to describe modern methods of oceanographical research, and thence to the large-scale features of the North Atlantic circulation, and the hydrography of the Gulf Stream. The reader is given an admirable account of the main characteristics of the Stream itself. Between leaving the Florida Straits and passing Cape Hatteras the Stream shows some of the characteristics of a jet. In its instantaneous form the spread of the actual Gulf Stream is less than 1°, and although its mass transport increases after leaving the Florida Straits, no further increase occurs after Cape Hatteras.

This account is followed by a treatment of fundamental problems concerning the Stream's existence including the geostrophic relationship and wind-driven ocean currents, leading on to discussions of the non-linear theories of the Stream proposed by various authors. Next, some attention is devoted to recent theories attempting to explain the problems peculiar to the Stream, such as the development of meanders and fluctuations in its transport and the breakdown of the Stream into currents and eddies north of Cape Hatteras, after which the role of salinity in oceanic circulations is considered in detail. The author shows that a qualitative explanation of the Gulf Stream system is possible on the assumption that it is wind-driven. Although the other theories he describes mostly serve to explain special features of the Stream such as the existence of counter currents, no all-embracing theory has yet been devised. The overall picture of the Gulf Stream which is drawn by the author is somewhat different from that existing in the popular imagination. It is that of a narrow intense current flowing along the western boundary of the warm Sargasso Sea which, by virtue of Coriolis forces, restrains the warm Sargasso water from overflowing the colder northern water of the North Atlantic. The Stream is not significantly different in temperature from the large mass of warm Sargasso water on its right.

Though some evidence is available that marked fluctuations in the mass transport of the Gulf Stream occur within periods of two months or less, the author admits that such studies are as yet in their infancy. Every oceanographic cruise so far made has found that the Stream and the immense store of available potential energy in the Sargasso Sea would suffice to maintain the density structure of the Gulf Stream for a long period in the absence of the stress of the winds. This largely accounts for the Stream's existence independent of the weather. But a paucity of data hinders all attempts to provide answers to questions of meteorological interest, as to how meteorological factors on any time scale influence the mass transport, breadth and depth of the Gulf Stream, or its geographical location and region of break-up; or how the Gulf Stream affects the airmasses flowing over it.

In a significant statement the author points out that the Gulf Stream itself is less likely to be important for the European climate than is the position and temperature of the large mass of water on its right flank, i.e., the surface waters of the Sargasso Sea. He cites some arguments by Iselin to the effect that a weakening of the surface water circulation in the North Atlantic (due to a previous weakening of the Azores-Bermuda sub-tropical anticyclone) should be accompanied by a rising thermocline and a spread of warm water radially outward and further northward than usual, thus warming the European climate, and then pleads for caution in drawing inferences regarding the effects of ocean currents upon existing climates in the present inadequate state of our knowledge. He goes on to deprecate use of the theory of wind-driven ocean currents by various workers to draw inferences relating to ocean-current systems and climates far back in geological time, where these were based upon distributions of land and ocean radically different from the present.

The author emphasizes the practical difficulties encountered on oceanographic cruises and pays tribute to the work of those who have overcome them. He describes promising field experiments carried out recently by aeroplanes over the Gulf Stream in which an infra-red bolometer was used, enabling rapid observations of sea-surface temperature to be made to an accuracy of 1°F. He makes vigorous pleas for better planned oceanographic cruising programmes in future, and for close repetitions of some measurements after an interval of years. Seasonal and secular changes in ocean-current features in the North Atlantic, which may have an economic significance for our fisheries, might thus be disclosed and related to meteorological factors.

The book is well produced and indexed and contains an extensive bibliography. It is perhaps unfortunate that the data sources quoted are nearly all from the U.S. and that the chapter on the wind system over the North Atlantic shows no evidence of collaboration with meteorologists on both sides of the Atlantic. The choice of synoptic situations in Figs. 49 and 50 together with their captions is peculiar, and the inclusion of the dates would have added to their value. However, these are minor blemishes in a book of this stature which could be profitably studied by any meteorologist who is concerned to gain fresh insight into the inter-dependence between oceanic and atmospheric processes.

R.F.M.H.

Particulate clouds, dusts, smokes and mists. By H. L. Green and W. R. Lane. Foreword by Sir Harold Hartley. London (E. & F. Spon). 1957. Pp. xix, 425; many Figures and Diagrams; 8 Plates. 70s.

In this book the authors present a systematic account of dusts, smokes and mists, incorporating the results of the vast amount of research that has taken place in the past three decades. In these important fields the authors are both well known and it is a testimony to their courage that they must have realized at the outset the immense task confronting them. The literature on particulate clouds is enormous, covering their physical and chemical properties, their behaviour in the atmosphere, their effects on health and many other aspects. In these subjects, perhaps more than in others, writers find great virtue in repetition and in the preface the

authors appeal very sensibly for a curb on the practice of publishing several papers in different places to describe one piece of work. However, even without such frustrations the authors undertook a mammoth job. There is no doubt it was a necessary job and few will dispute that it has been very well done.

The first part of the book deals with the basic scientific aspects with chapters on the production of particulate clouds, on their physical characteristics, optical properties, coagulation, deposition and sampling. The final chapter in this section of the book is concerned with the diffusion of suspended matter in the atmosphere and, within the space available, provides as good an introduction to the subject as can be found anywhere.

Dusts, smokes and mists are nowadays inescapable features of our daily lives and in the second part of the book the authors treat of the industrial and environmental aspects. It begins with a chapter on the various techniques of collection; then a chapter on health hazards is followed by one on atmospheric pollution and one on the natural occurrence of aerosols. The final chapter contains a description of the various uses of particulates in industry, in agriculture and in pest control, and for therapeutic purposes.

I am confident that this book will prove of great value to workers with general interests in this wide field but it should also be useful to the specialist – not so much in his own subject but for the ready access it provides to up-to-date information about neighbouring or overlapping subjects. Newcomers to the study of particulates and allied topics will find the book most useful for its comprehensive survey and for its guide to more detailed reading.

The writing and the production are of a high standard throughout. The diagrams and plates are very good and the bibliographies appended to each chapter appear to have been drawn up with the greatest care.

P.J.M.

Annual Report on the Meteorological Office for the year ending 31 March 1958, (M.O. 650). London (H.M. Stationery Office). Pp. 52; Tables; 4 Plates. 3s. 6d.

Much of the Director-General's foreword to this report is naturally concerned with the outstanding event of the Meteorological Office year – the reorganization in July 1957 of the Office, following the report to the Secretary of State for Air of the Brabazon Committee.

Some of the changes now in force are far-reaching and important. The creation of more high-level posts brings a needed improvement in the career prospects of both the scientific and experimental officer classes. It is, however, a matter of some regret that the improvement for the scientific officer class should rest, even in small part, on a reduction of the overall establishment of this class. The impetus given to research is particularly marked and is well justified by the high quality of the research work done in the Meteorological Office in recent years. A novel and imaginative step is the creation of a post in which the main function will be to ensure a speedy two-way traffic between the research worker and bench forecaster.

That section of the report written by the Director of Services shows that the work of the Meteorological Office touches the national life at almost every point, and that the resources of the Office are continuously strained to meet its commitments. There was a small net decrease, both at home and overseas, in the number of stations operated for the military services : in almost every other field the demands made on the office continue to grow. Increasing use of modern methods of data processing, an experimental decrease in the documentation provided for civil aviation flights, and the application of Work Study techniques are examples of the manner in which this problem is being met. Outstanding advances in communications included the partial replacement of wireless telegraphy by higher-speed radio-teleprinter broadcasts and an experimental two-way transmission of facsimile weather charts between the United Kingdom and Canada.

The Director of Research briefly discusses the special position and responsibility of the Meteorological Office with respect to the prosecution of meteorological research in the United Kingdom and summarizes the current position of Meteorological Office research in dynamical and other forecasting problems, climatology, atmospheric physics and instrument development. The imminent installation at Dunstable of an electronic computer was eagerly awaited.

The report continues with a discussion of the participation of the Office and some of its staff in the field of international cooperation in meteorology and geophysics, including the International Geophysical Year, and ends with a summary of data relating to the organization and administration of the Office. It is an impressive document which leaves no doubt that the many functions of the Meteorological Office are being performed in a most capable and vigorous manner.

D.H.M.

Air pollution : Film strips (in colour) and Lecture notes. (A) Nature, production and effects of air pollution. (B) Dispersion and accumulation of air pollution : behaviour and appearance of smoke. By R. S. Scorer. London (Diana Wyllie), 1959. 32 frames per strip, 35 × 24 mm. Lecture notes 29 pp. £4 complete.

In a foreword to this collection of colour filmstrips Sir Hugh Beaver says that they will make a real contribution to the study of the many problems of air pollution. That is an opinion that few will quarrel with for Dr. Scorer has provided a rather spectacular survey of the whole field, laying emphasis wherever possible on the scientific aspects.

Section A deals with the nature (grit, smoke, gases, etc.), the sources (industrial plants, domestic fires, road vehicles, etc.) and the effects (on health, vegetation, buildings) of air pollution. There is also a warning that unless adequate precautions are taken nuclear power stations may inject into the atmosphere a deadlier form of contamination than the sulphur and smoke which have caused so much harm in the past.

The dispersion and accumulation of pollution are the subjects of Section B and, as one would expect, the treatment is largely meteorological, but in a manner that should be welcomed by those who specialize in other fields of air pollution without quite being able to shuffle off the mortal coil of meteorology.

The filmstrips and lecture notes must be judged principally on the educational objective which the author had in mind and here we should distinguish between two aspects – the teacher and the taught. For the latter, Dr. Scorer has provided the material for a most interesting introductory course in air pollution and one could wish that lectures using this material could be given up and down the country under the patronage of local authorities who now have important responsibilities under the Clean Air Act. The audiences, after a stimulating hour, would repair to their homes convinced that air pollution is bad, is excessive and should be reduced – and not much more. Such is the way with lectures, but the right frame of mind would have been created.

As regards the lecturers, it must be remembered that lecturers as a class are regarded as experts and it is important they should form balanced judgements on difficult and sometimes insoluble problems. Lecturers who have recourse to these film strips and notes will find their work done for them, and most competently done. And yet, it seems a pity that Dr. Scorer in his notes should guide the lecturers, who may not all be critically minded, to form decisive opinions on such questions as gas washing and the apportioning of blame among the various sources of pollution, about which a great deal still remains to be learned. It is right that these problems should be discussed, but a lecturer should speak from a wide understanding and not a restricted one. I regard this blemish in an otherwise excellent piece of work as important because great value is placed on the question period which normally follows a lecture.

It would, however, be unjust to end this review on a critical note which applies only to a small fraction of the whole work. The selection and production of the film strips have been admirably done and the lecture notes have the clarity of expression for which the author is well known.

P.J.M.

THE NAPIER SHAW MEMORIAL PRIZE

Announcement of third competition

The Royal Meteorological Society announces the third competition for the Napier Shaw Prize. The competition will be open to anyone without restriction of nationality. The prize of £100 is offered on this occasion for an original essay on '**The Upper Atmosphere.**'

> For the purpose of this competition the upper atmosphere is regarded as the atmosphere above the tropopause and below the exosphere. Essays on the dynamics, constitution or physics of the atmosphere in this region, or in any part of it, will be accepted, but an essay on purely electrical properties of the ionosphere is not sought. Preference will be given to essays relating two or more aspects of the state of the upper atmosphere. Essays on the inter-relations between the upper atmosphere and the troposphere will also be welcome.

The essay, which must be in English, may include material which the candidate has already published. Essays may be submitted jointly in the names of more than one author.

Candidates are required to submit three copies of their essay, which must be printed or typewritten. In the latter case, only one side of the paper should be used, double spacing should be employed and two of the copies may be good carbon copies. Unless the essay is printed, any mathematics should be written in carefully by hand. Although the scale of the essay will be considerably greater than that of an ordinary paper, candidates are recommended to consult, for general advice, the pamphlet, 'Preparation of papers for the Quarterly Journal.' This may be obtained, free of charge, on application to the Assistant Secretary.

It is required that any essay for which a prize is awarded shall be offered to the Society for publication.

All entries for the competition should be addressed to the Assistant Secretary, Royal Meteorological Society, 49 Cromwell Road, London, S.W.7 and marked clearly 'Napier Shaw Prize.'

Essays for the competition must be received at the Society's Offices *not later* than 1 January 1962.

Obituary Notices

Mr. E. SATTERTHWAITE

Mr. Edwin Satterthwaite, whose death occurred in January 1959 at the age of 47, was a native of Ulverston, Lancashire, where he spent the greater part of his life. He held the post of chief clerk to the Barrow-in-Furness County Court, having worked in that office since the year 1928, except during the war period, when, after transfer to Workington as clerk-in-charge in 1939, he entered the R.A.F. in October 1941, serving as a radio mechanic until 1945. Mr. Satterthwaite had studied meteorology from boyhood, and his rainfall records go back to 1927. His unpublished papers include a detailed account of the severe winter in 1947, illustrated with many maps and charts. He contributed letters concerning the weather to local newspapers. His friends will remember him as a meticulous and devoted student of meteorology, with a true love of the subject that engrossed so much of his leisure-time. Mr. Satterthwaite was elected a Fellow of this Society in 1950.

E.S.

Mr. D. HALTON THOMSON, O.B.E., M.A., M.I.C.E.

David Halton Thomson, a Fellow of the Society since 15 March 1933, died at the age of 73 on 17 January 1959 after a short illness. Having graduated from Cambridge with honours in the Mechanical Sciences Tripos, Mr. Thomson joined the firm of Messrs. Worthington-Simpson, pumping-engine manufacturers, Newark-on-Trent. From 1912 he served with the Borough of Portsmouth Waterworks Company (now the Portsmouth and Gosport Water Company), initially as chief engineering assistant and latterly, from 1926, as engineer to the Company – a post which he continued to hold until his retirement in 1950. During his tenure of office in this capacity he carried out many important measures for ameliorating the provision of water to the Portsmouth district. At the centenary of the Company in 1957 Mr. Thomson wrote *A History of the Portsmouth and Gosport Water Supply*, in which he was able to trace the record back to the reign of Henry VIII.

Mr. Thomson was a past-president of the Institution of Water Engineers and at the time of his death held office as its honorary secretary. He was also a former president of the Portsmouth Chamber of Commerce. His keen interest in rainfall and climatology together with mathematical and statistical ability made him well qualified to undertake the close study of hydrological conditions in the chalk of the South Downs which, through a series of able papers, won him recognition as a leading authority on the subject. At a meeting of this Society in November 1954 he was the opening contributor to a discussion on the water problem in Great Britain. Mr. Thomson received the award of the O.B.E. for Civil Defence work during the 1939-45 war.

H.O.P.
E.L.H.

Mr. G. MICHELL, F.R.C.O.

Among the many Fellows of this Society whose interests include music as well as meteorology was Guy Michell, who died at his home in Hove, Sussex, on 6 July 1959 towards the close of his 83rd year. As an organist and composer of repute, Mr. Michell had played a prominent part in the musical life of Sussex and the Isle of Wight for well over half a century. Concerts of his compositions were also given elsewhere in Great Britain and in the United States. A barometer received as a present on his 16th birthday led him to take up the study of weather science with an enthusiasm which proved to be abiding. Thereafter observational meteorology had a constant claim on his attention, and from about 1930 he contributed data to the official annual publication *British Rainfall*. Mr. Michell had achieved a considerable local reputation as a weather forecaster. His Fellowship in the Society dated from 19 February 1930.

E.L.H.

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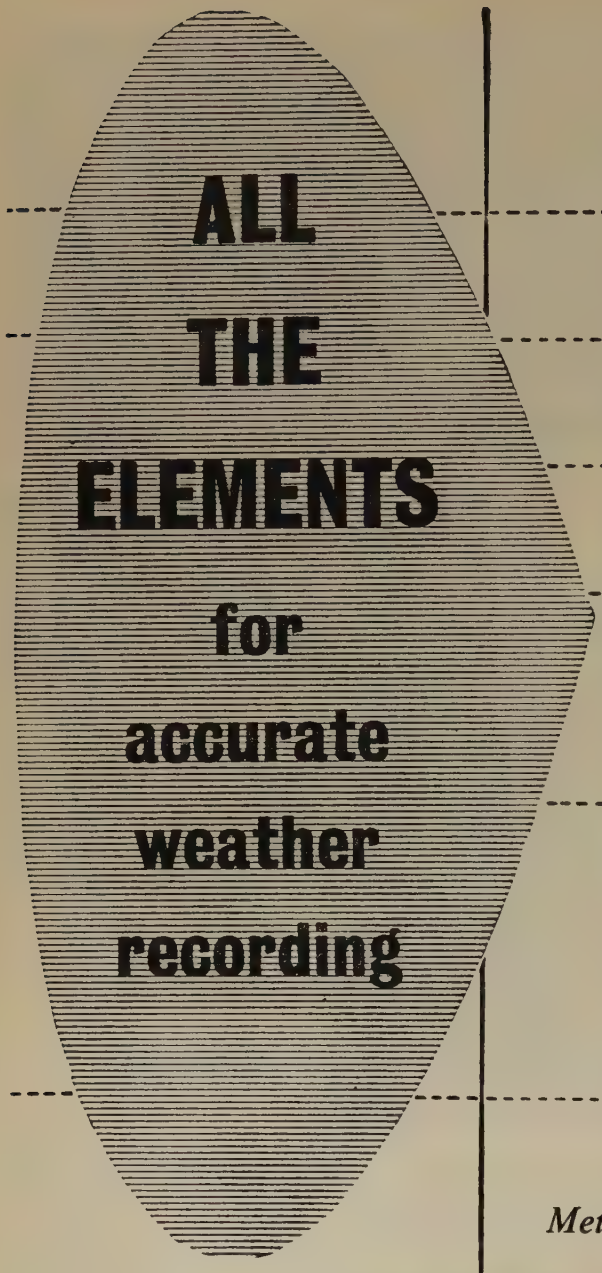

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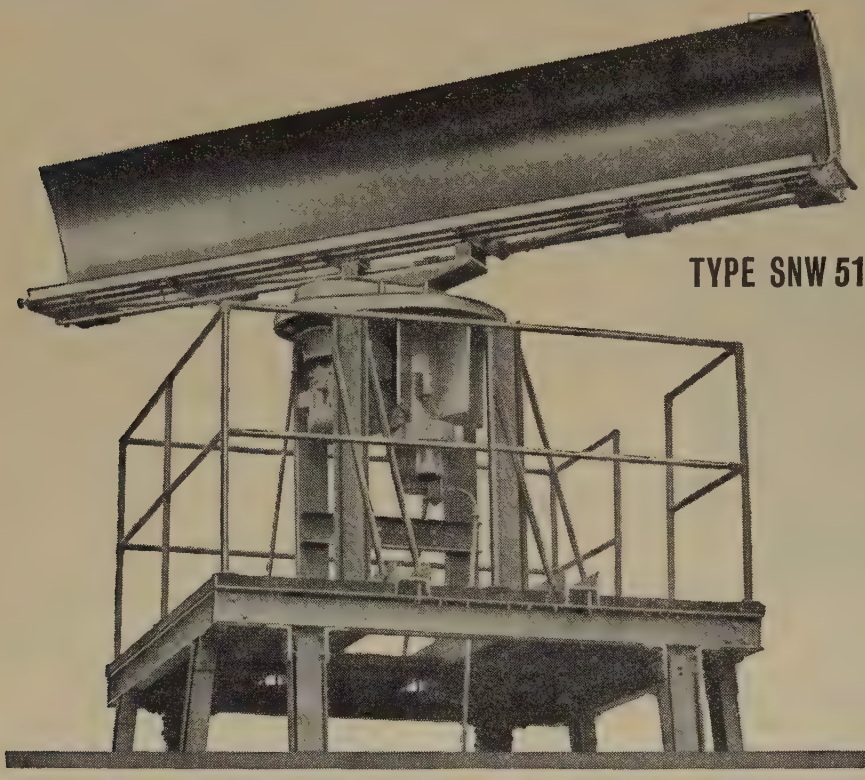

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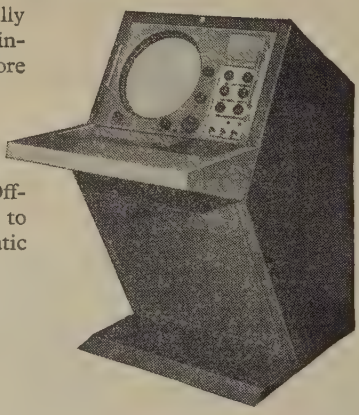

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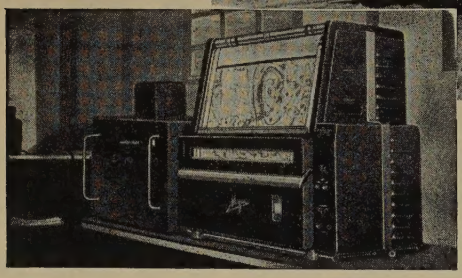

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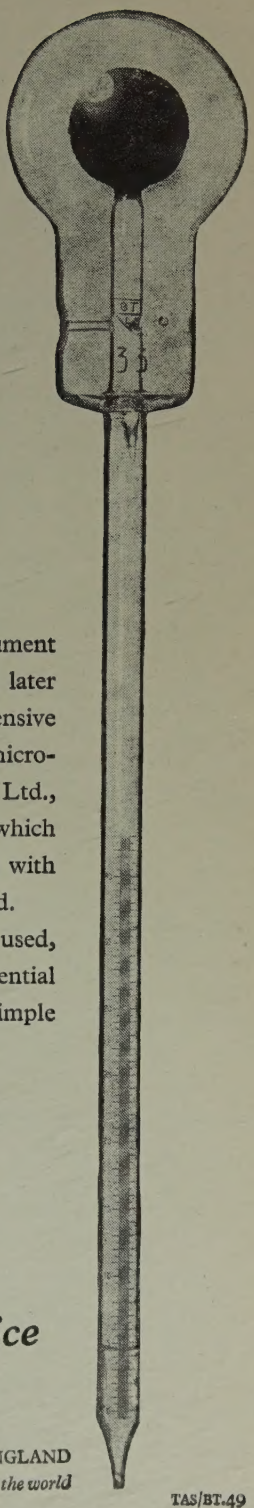

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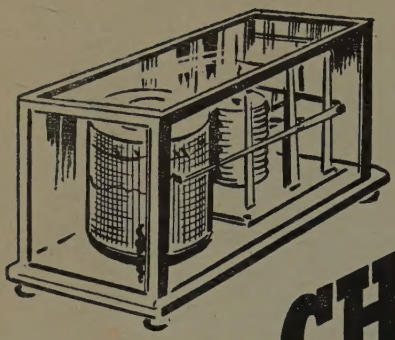

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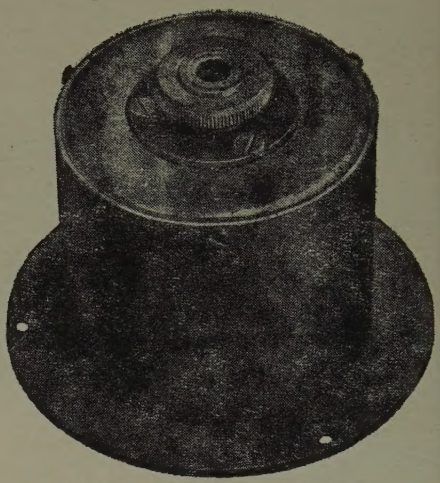

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